Volume 2, Number 6 November-December 1994

ISSN: 0869-5938

STRATIGRAPHY AND GEOLOGICAL CORRELATION

Official English Translation of *Stratigrafiya, Geologicheskaya Korrelyatsiya*

Editor-in-Chief
Boris S. Sokolov

Academician, Russian Academy of Sciences
Counsellor of the Presidium of the Russian Academy of Sciences

Journal of the Fundamental and Applied Aspects of Stratigraphy and the Correlation of Geologic Events and Processes in Time and Space

МАИК НАУКА/INTERPERIODICA PUBLISHING

Stratigraphy and Geological Correlation

(Stratigrafiya. Geologicheskaya Korrelyatsiya)

ISSN: 0869-5938

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Stratigraphy and Geological Correlation is indexed in *Petroleum Abstracts* and *World Translations Index.*

SUBSCRIPTION OFFICE
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P.O. BOX 1831, BIRMINGHAM, AL 35201-1831 USA
PHONE: (800) 633-4931; (205) 995-1567 (outside US and Canada); FAX: (205) 995-1588

PUBLISHER
МАИК НАУКА/INTERPERIODICA PUBLISHING
PROFSOYUZNAYA UL. 90, MOSCOW, 117864 RUSSIA
PHONE: 7-(095) 336-0066; FAX: 7-(095) 336-0666

SUBSCRIPTION PRICES (1994, Volume 2, 6 issues)

US $591.00 (North America), $645.00 (other). A 5% discount is given on 2-year orders. Subscriptions may be entered at any time. Orders may be placed at agencies or directly. 5% discount to all subscription agencies. Cancellations not allowed except for duplicate payment. Prepayment required. Advance payment required in US$/US bank. MasterCard/Visa accepted. Checks payable to *Stratigraphy and Geological Correlation* or Interperiodica.

Stratigraphy and Geological Correlation (ISSN: 0869-5938) is published bimonthly for (US dollars) $591.00 (North America), outside North America $645.00 per year by MAIK Nauka/Interperiodica Publishing, 5724 Hwy. 280 East, Birmingham, Alabama 35242-5724 USA. Application to mail at Second-class rates is pending at Birmingham, Alabama. POSTMASTER: Send address changes to *Stratigraphy and Geological Correlation*, P.O. Box 1831, Birmingham, Alabama 35201 USA.

Journal founded in 1993.

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Some General Features of Mammalian Evolution

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Received October 27, 1993

Abstract – Study of Neogene and Quaternary mammals showed that the evolution of members of single phyletic lineages in a number of mammalian groups follows a general law that can be described by an inverse hyperbolic cotangent formula. This law is applicable when solving a number of theoretical and practical problems, such as the unification of the concept of biological and paleontological species, the determination of morphological boundaries between species and their life spans, and the determination of evolutionary rates. Another application is an elaboration of the biometrical method in the precise dating of bone remains.

Some mammalian phyletic lineages show directional evolution, which is expressed in time-dependent changes of morphology. Analysis of this data revealed that some of these changes may develop according to a common law. We came to this conclusion on the basis of both original data for genera *Promimomys*, *Borsodia*, and *Archidiskodon–Mammuthus*, and data from the literature for genera *Elephas* and *Mesochoerus* (Cook and Maglio, 1972), *Dicrostonyx* (Kochev, 1989), and *Pliopotamus–Ondatra* (Martin, 1984) (Tables 1 - 3).

For the genus *Promimomys* (time span of 7.83 to 3.58 Ma), as a morphological evolutionary parameter (MEP) we used a sum of the heights of the three dentine tracts of the first lower molar. For the genus *Borsodia* (2.61 - 2.25 Ma), the height of the inner posterior tract in the first lower molar was used. In the case of the *Archidiskodon–Mammuthus* lineage (2.35 - 0.64 Ma), the number of plates within 10 cm intervals of the third upper molar was used. In the case of the genus *Elephas* (4.1 - 0.04 Ma), we took a height/width ratio in the third upper molar as an MEP; in genus *Mesochoerus* (3.00 - 1.00 Ma), the length of the third upper molar. For *Dicrostonyx* (0.225 Ma – recent), we used the depth of additional synclines in the first and second upper molars, which is expressed through the value of the angle inscribed into these synclines. Finally, for the lineage *Pliopotamys–Ondatra* (3.98 – present), MEP is the height of the first labial tract in the first lower molar.

In all cases, time-dependence of the value of the morphological evolutionary parameter can be described by the formula of an inverse hyperbolic cotangent, with $x > 1$. An inverse hyperbolic cotangent has a logarithmic expression according to the formula:

$$y = \frac{1}{2}\ln\frac{x+1}{x-1}, \qquad (1)$$

where x is geological age of the studied object, and y is the value of MEP.

From formula (1) it follows that:

$$x = \frac{e^{2y}+1}{e^{2y}-1}. \qquad (2)$$

The positions of the phyletic lineages considered in the curve of an inverse hyperbolic cotangent are shown in Fig. 1.

The establishment of this law enables us to deal with a number of practical and theoretical problems.

BIOMETRIC DATING METHOD

Knowing the law that describes the value of MEP as a function of time, we can also solve the inverse task of determining the age of fossil localities on the basis of MEP.[1] This can be done either in graphical form or from the formula (2). The latter method is by far more precise. In both cases, however, scales should be chosen for axes X (time) and Y (morphological evolutionary parameter). At least three points (in our case, three localities) with known values of MEP and age are required to accomplish this. Then the position of the three points system (keeping constant the relationships of time and MEP between the first and second and the second and third points) is shifted along the inverse hyperbolic cotangent curve. By changing the scale along axes X and Y, we can find their position on the curve where the time dependence of MEP, which is immanent for the system, is maintained.

As an example, let us consider the age determination of localities with *Promimomys* remains from Moldova and southern Ukraine. The Musait 5, Budei, Etuliya (lower and middle fossiliferous beds), and Lucheshty VII localities are from the Karboliya fluviatile deposits in western Moldova.

[1] The idea of using biometric data for determining the age of localities has been suggested by Martin (1984) and exemplified by an empirical curve constructed for the lineage *Pliopotamys–Ondatra*.

The Grebeniki 2 (southern Ukraine) and Novaya Andriashevka (eastern Moldova) localities are confined to deposits of the Kuchurgan Gravel. The Mugureny locality (central Moldova) is from Stol'nicheny fluviatile deposits, and the Vinogradovka locality (Bolgrad district of Ukraine) is associated with lower beds of the lower Pontian deposits (Vangengeim *et al.*, in press).

In this case, MEP is a sum of the heights of the three dentine tracts in the first lower molar. These tracts are the outer tract of the anteroconid and the outer and inner tracts of the posterior prism. The height of the dentine tracts was measured from the conventional line drawn at the base of the crown, normal to the long axes of synclinal folds, through the lowermost point of enamel in the posterior external synclinal fold.

In order to determine the position on the inverse hyperbolic cotangent curve and the scales along the X and Y axes, we took localities Vinogradovka, Grebeniki 2, and the lower horizon of the Etuliya locality as three reference points. The fossiliferous bed of the Vinogradovka locality lies in reversely magnetized sediments of the lower Pontian (magnetic Chron 6), immediately above a zone of normal polarity (Chron 7). On the basis of the age of the boundary between Chrons 6 and 7, which is 6.8 Ma, the age of the fossiliferous bed is assumed to be 6.75 Ma. The age of the Grebeniki 2 locality is taken as 3.95 Ma. This age is based on the fact that *Promimomys moldavicus* from the Grebeniki 2 locality is evolutionally identical with *P. moldavicus* from the Malusteni locality of Romania. The age of the third site is estimated to be from 4.0 to 3.9 Ma (Radulescu and Samson, 1989).

The age of localities from the Karboliya fluviatile deposits is restricted from below by the age of the top of the Kochiti Subchron, and from above by the age of the Gilbert–Gauss boundary.The geologic position of the lower bed of the Etuliya locality within Karboliya fluviatile deposits suggests a correlation of the site with the middle part of the reversely magnetized interval between the top of the Kochiti Subchron and the Gilbert–Gauss boundary. Hence, its age is assumed to be 3.60 Ma.

From formula (2), after determining the position of the reference points on the curve and the scales along the axes, we obtain the formula for calculation:

$$T = 3.5\frac{2.6e^{1.9M} + 1}{2.6e^{1.9M} - 1} \text{ Ma.} \qquad (3)$$

Ages of the localities, calculated according to (3), are given in Table 3.

As is evident from the Table, absolute (k) and relative (k') errors of age determination are not equal. They range from 0.005 to 0.163 Ma (k), and from 0.14 to 4.12% (k'), respectively. This difference in accuracy is fairly explicable. An error in age determination depends mainly on the position of the mean value of MEP (M) on the inverse hyperbolic cotangent curve and the precise determination of M. Error in the determination of M depends on the value of root-mean-square deviation from $M(\sigma)$ and the number of measurements (n). The value of σ is independent because any parameters of variation are an immanent characteristic of the data; thus, the accuracy of M determination can only be improved with an increase in the number of measurements. In this case, the mean error can be reduced as strongly as is wished, and the accuracy of M determination can be raised to any desired value.

In actual practice, it is natural that the number of measurements cannot be increased infinitely. This is particularly true for paleontological material. Still, age determination error not exceeding 1% can already be reached with a relatively low number of observations.

Highly accurate age determination for the localities of forms from the same phyletic lineage enables the determination of time intervals between these localities with a minor error that is less than the first several thousands of years. However, the whole system can be shifted up or down the geochronological scale according to the value of error of the absolute age of the reference localities.

Thus, we cannot increase the absolute age of all localities biometrically dated on the basis of *Promimomys* remains more than 35 ka. This is because the Vinogradovka reference locality, situated in reversely magnetized deposits (beginning of Chron 6), cannot be older than the boundary of magnetic Chrons 7 and 6, which has an age of 6.8 Ma. Similarly, the age of all sites cannot be decreased more than 90 ka. This conclusion is substantiated by the fact that *Promimomys* from the Grebeniki 2 locality (3.95 Ma) is older than the vole from the Lupoaia locality in Romania. According to Andreescu and others (Andreescu *et al.*, 1986), the age of the latter locality is slightly older than that of the top of the Kochiti Subchron (3.86 Ma).

SOME CONSIDERATIONS ON PALEONTOLOGICAL SPECIES LIMITS

The problem of species in paleontology is extensively addressed in the literature but until now remains unsolved. It is well-known that the criteria of interbreed among individuals of the same species and reproductive isolation among those of different species are based on the neospecies concept. In practice, however, neontologists frequently use only the morphological criterion of species (Cain, 1954; Mayr, 1970; Timofeeff-Ressovsky *et al.*, 1977; and others).

In paleontology, genetic criteria (interbreed and reproductive isolation) are inapplicable for obvious reasons. Therefore, a concept of the paleospecies and the neospecies arose. Cain (1954) was of the opinion that the only major difference between neo- and paleospecies is the natural morphological limits of the former and the arbitrary limits of the latter.

In the case of a stratigraphic gap in the fossil record, a morphological break between two forms provides a basis for classifying them as separate species. Yet even in this case, there is no certainty that they are not simply

extreme forms within the variation range of a single species. In the case of continuous phyletic evolution, it is even more difficult to draw a boundary between species.

Gromova (1959) thought that it is necessary "... to achieve the greatest possible similarity between the volume and content of fossil species and the species concept in modern biology, and in neontological systematics in particular" (p. 13). Simpson (1961) believed that morphological distances between fossil species should be at least no smaller than the distances between recent species of the same or related taxonomic groups.

We share this opinion and propose one of possible approaches to the problem. This approach enables us to use morphological criteria to bridge the gap between concepts of volume and content of fossil and recent mammalian species.

We propose to take extreme values of the variability range of an MEP of a recent species, expressed as percentages, as the morphological limits of a fossil species. In doing so, we consider the variability to be constant in time.

The proposed approach can be exemplified with American muskrats. The height of the first labial tract of the first lower molar is used here. Let the distance from the X axis to point E (recent species *Ondatra zibethicus*) be 100% (Fig. 2). In this case, the maximum value of variation of the dentine tract's height in recent species is ±16% of the distance between the X axis and point E.[2] Hence, the lower limit of the species amounts to 84% of the distance between the X axis and the mean value of MEP, and the upper limit equals 116%. It is assumed that these values are valid for all fossil species of the phylum. Correspondingly, we can calculate values of MEP for species limits.

Zero value of MEP is taken as the lower limit of the first species of the lineage. The distance between the X axis and the point with a zero value of MEP will amount to 84% of the distance from the X axis to the mean value of this species. The upper limit of the first species is, accordingly, the lower limit of the next species. Distance from the X axis to this limit will equal 116% of the distance from the X axis to the mean value of the first, 84% for the second species, and so on. Consecutive calculations for the whole set gave positions for the boundaries of all species of the lineage, expressed in MEP units. They are marked with dashes in the left part of the plot. Values of MEP at the species boundaries correspondingly provide an opportunity to calculate the ages of these boundaries and the life spans of the species.

Our calculations show a presence of 10 species in the considered lineage (I - X, Fig. 2). The first species (I) is *P. minor*. Martin (1984) pointed out that tracts of this form in the Hagerman locality fauna are very

[2] Values of variation and means of height of the dentine tract are taken after Martin (1984). If more detailed studies change these values, the corresponding calculations will be modified without sacrificing the main idea.

Table 1. Mean values of morphological evolutionary parameter and ages of localities with remains of *Archidiskodon–Mammuthus, Borsodia, Pliopotamys–Ondatra*

Localities	T	n	M	T^1	
Archidiskodon–Mammuthus[1]					
Tiraspol'	0.64*	19	6.1	0.643	+0.074 −0.062
Sinyaya Balka	0.95*	45	5.50	0.947	+0.061 −0.054
Liventsovka, lower bed	2.35*	24	4.58	2.355	+0.164 −0.149
Borsodia					
Lebyazh'e	2.32*	10	2.77	2.325	±0.007
Kryzhanovka 1, 2	2.43*	9	1.77	2.427	+0.016 −0.017
Veselovka	2.53*	5	0.94	2.530	+0.022 −0.021
Pliopotamys–Ondatra[2]					
Recent	0*		6.1	0.000	
Visconsin	>0.1		5.2	0.013	
Cudahy	0.6*		1.08	0.603	
Borchers, Grandview	1.9*		0.29	1.843	
Hagerman	3.5		0	3.982	

Note: T – age in Ma based on geologic, radiometric, or paleomagnetic methods; * – reference points; n – number of samples; M – mean value of morphological evolutionary parameter; T^1 – age in Ma, calculated by biometric method; [1] values of M and n are after Dubrovo (1964, 1975), Dubrovo and Baigusheva (1964). [2] values T and M are taken from the plot (Martin, 1984, Fig. 7).

Table 2. Mean values of morphological evolutionary parameter and ages of *Elephas, Mesochoerus*, and *Dicrostonyx-species*

Species	T	n	M	T^1
Elephas[1]				
E. iolensis	0.04*	8	2.37	0.040
E. recki, stage 3 - 4	1.0*	10	1.71	0.988
E. recki, stage 1 - 2	2.2	13	1.39	2.182
E. ekorensis	4.1*	6	1.15	4.100
Mesochoerus[2]				
M. olduvaiensis	1.0*	5	61.5	1.0
M. limnetes	1.9*	8	48.1	1.895
"	2.1	8	45.9	2.095
"	2.45	7	42.2	2.456
"	3.0*	4	36.7	3.094
Dicrostonyx[3]				
D. torquatus	0*		74.7	0.000
D. gulielmi gulielmi	30*		37.6	30.047
"	37		32.6	37.040
D. simplicior posterior	210		1.4	204.837
D. simplicior simplicior	225*		0	225.471

Note: Legend is the same as in Table 1. T and T^1 for *Dicrostonyx* are in ka: [1] and [2] values of T and M are taken from plots (Cook and Maglio, 1972, Figs. 7 D and 5, respectively). [2] values of T and M are taken from the plot (Kochev, 1989, Fig. 1).

Table 3. Age of mammalian localities calculated biometrically on the basis of the *Promimomys* genus

Localities	n	M	m	σ	T	k	k'
Etuliya (middle bed)	10	1.84	±0.03	±0.11	3.583	±0.005	±0.14
Lucheshty VII	18	1.79	±0.04	±0.18	3.591	±0.007	±0.19
Etuliya (lower bed)	9	1.69	±0.05	±0.14	3.610	+0.011 −0.008	+0.30 −0.22
Budei	2	1.25	±0.05	±0.07	3.760	+0.027 −0.024	+0.72 −0.63
Musait 5	5	1.18	±0.08	±0.17	3.798	+0.051 −0.044	+1.34 −1.16
Grebeniki 2	3	0.97	±0.15	±0.25	3.954	+0.163 −0.118	+4.12 −2.98
Novaya Andriashevka	5	0.62	±0.03	±0.03	4.441	+0.062 −0.059	+1.40 −1.33
Mugureny	1	0.23			5.814		
Vinogradovka	2	0.10	0	0	6.765		

Note: n – number of remains (first lower molars – M_1); M – mean value of the sum of dentine tracts in mm; m – standard error of M in mm; σ – standard deviation from M value in mm; T – age, Ma; k – standard error of age determination in Ma; k' – relative error of age determination in percents.

rare or absent. He took the values of this fauna as the zero value of MEP. But, as is clear from our calculations, the zero value of MEP should be peculiar to a form with an age of 3.982 Ma. An age of 3.5 Ma (the date for the Hagerman locality) implies a mean value of 0.037 mm for the dentate tract. The second (II) and third (III) species are *P. meadensis* and *Ondatra idahoensis*, respectively. The fourth species (IV) represents a form of Arvicolid Zone IV and was unnamed by Martin. The fifth species(V) is *O. annectens*. A form that corresponds to species VI is so far unknown. The seventh species (VII) is *O. nebraskensis*. Species VIII and IX may correspond to forms from Mount Scott and the Millers Cave locality (Nelson and Semken, 1970). The last species of the lineage (X) is *O. zibethicus*. According to our calculations, this recent form appeared 13.5 ka ago.

The plot indicates that the species differ in their life spans, which decrease up the time scale.

In the given example, we determined specific limits of fossil forms on the basis of the variability of recent species. If a phyletic lineage is entirely represented by fossil species, the same approach for determining their limits remains appropriate. In this case, calculations are based on the variability of a particular feature of the most completely represented species from one locality or from several strictly synchronous sites. To illustrate this, let us use genus *Promimomys* from the southern Ukraine and Moldova localities (Fig. 3). The bulk of *Promimomys* remains is known from the Lucheshty VII locality. As an MEP, we take the sum of the heights of the three dentine tracts in the first lower molar. Variation of MEP is ±13% of the distance from the X axis to the mean value of MEP in the plot. Therefore, limits of the species amount to 87 and 113% of the distance from the X axis to the mean value of MEP for each species, respectively.

Zero value of MEP, as in the example of muskrats, is taken as the lower limit of the first species of the lineage. After calculations such as those described above, we obtain MEP values for species limits, marked by the dashes in the left part of the plot. Arrows in the time scale mark the age of these limits.

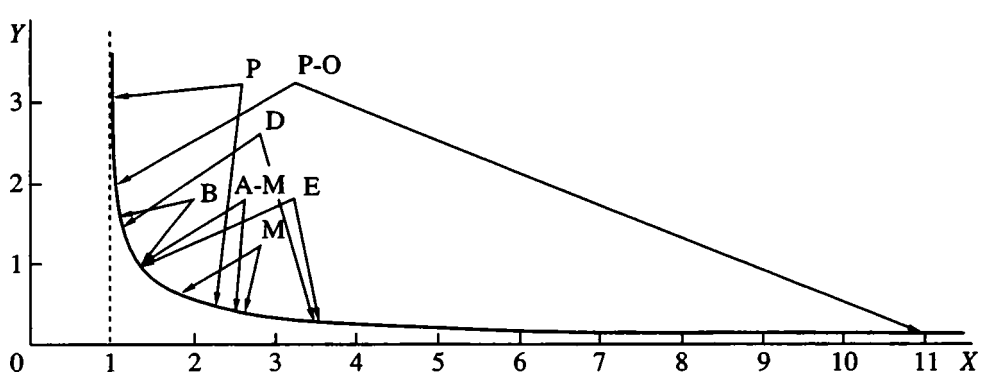

Fig. 1. Position of various phyletic lineages of mammals on the inverse hyperbolic cotangent curve constructed in conventional units. P – *Promimomys*; P-O – *Pliopotamys–Ondatra*; D – *Dicrostonyx*; M – *Mesochoerus*; B – *Borsodia*; E – *Elephas*; A-M – *Archidiskodon–Mammuthus*.

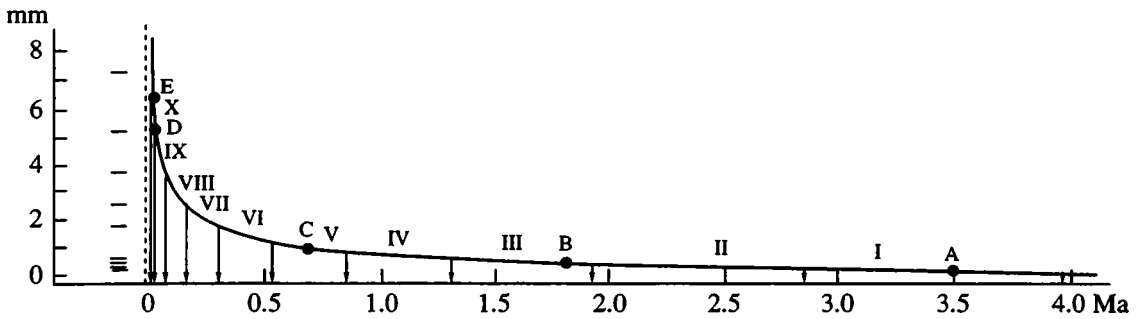

Fig. 2. Time-dependent changes of first labial tract of M_1 in the lineage *Pliopotays–Ondatra1*. *X* axis – time in Ma; Arrows indicate age of proposed limits of species (species I - X); *Y* axis: figures show height of first labial tract of M_1 in mm; horizontal dashes to the right of the *Y* axis indicate values of this parameter for the species limits. Localities: A – Hagerman; B – Borchers and Grandview; C – Cudahy; D – Visconsin; E – recent (after Nelson and Semken, 1970; Martin, 1984).

Two of the most ancient and still undescribed forms of *Promimomys* are represented by scanty remains from the Vinogradovka and Mugureny localities. The next, younger species is absent in our collection. The fourth species of the phyletic lineage is *Promimomys insuliferus* from the Novaya Andriashevka locality. The fifth species, *P. moldavicus*, is known from the Grebeniki 2, Musait, and Budei localities. It should be pointed out that the form from the Grebeniki 2 locality, as well as the specimen type of this species described from the Malusteni locality in Romania, are at the lower (older) limit of the species.

Finally, the last form, which is assigned to *P. moldavicus* by some authors, is known from the Etuliya and Lucheshty VII localities. This form should be considered a separate and terminal species in the lineage of genus *Promimomys*.

Thus, the given data indicates a continuous evolutionary lineage. In one way or another, all but one of its members are represented in the fossil material.

EVOLUTIONARY RATE AND STAGE OF EVOLUTIONARY ADVANCEMENT

In the literature, there is no sharp distinction between the concepts of tempo and rate of evolution. They are often considered to be synonymous. Timofeeff-Ressovsky *et al.* (1977, p. 203) thought that "In the determination of the tempo of evolution, two major approaches are possible: first, determination of the rate of particular character changes in the phylogenesis of a group; second, determination of the rate of new taxa origination."

In all phyletic lineages considered, time-dependent changes of separate features are recognized. As these changes are correlated with the time scale, a rate of feature change, i.e., the value of a feature change per time unit, can be calculated.

All phyletic lineages lack a uniform morphological feature. Because of this, the rate of change of separate features cannot be used for a comparison of evolutionary rates in different lineages. For this purpose, it is more logical to use the rate of origination of new taxa, i.e., the number of taxa originating per time unit.

Let us illustrate this by the example of the genus *Promimomys* and the *Pliopotamys–Ondatra* lineage. Mean rate of speciation in genus *Promimomys* is 0.0014 species per thousand years (6 species in 4260 ka). In the lineage *Pliopotamys–Ondatra*, the rate is 0.0025 species per thousand years (ten species in 3980 ka). It is clear that the rate of speciation in muskrats is 1.78-fold higher than in the genus *Promimomys*. It should be

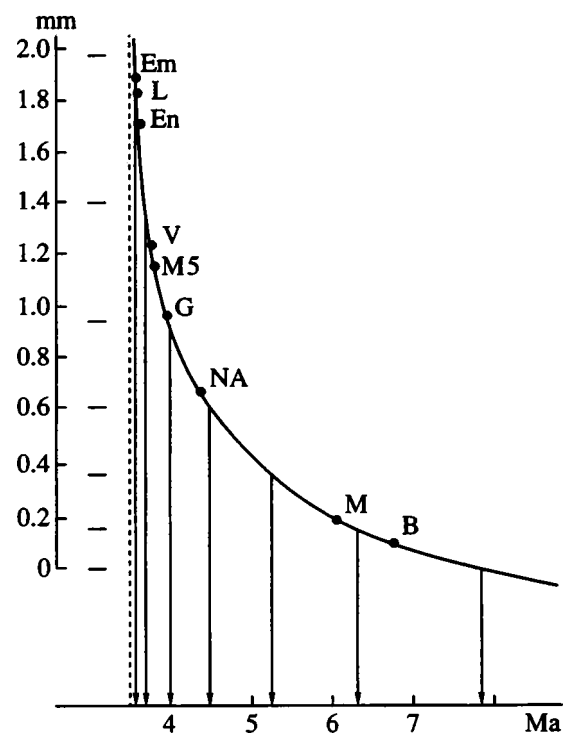

Fig. 3. Time-dependent change of the sum of the heights of three tracts of M_1 in species of genus *Promimomys*. *X* axis – time in Ma; arrows indicate age of proposed limits of species; *Y* axis: figures show the sum of the heights of three tracts of M_1 in mm; horizontal dashes to the right of the M_1 axis indicate values of this parameter for the species limits. V – *Promimomys* sp. n. 1, Vinogradovka; M – *P.* sp. n. 2, Mugureny; NA – *P. insuliferus*, Novaya Andriashevka; G – *P. moldavicus*, Grebeniki 2; M5 – *P. moldavicus*, Musait 5; B – *P. moldavicus*, Budei; En – *P.* sp. n. 3, Etuliya (lower bed); L – *P.* sp. n. 3, Lucheshty VII; Em – *P.* sp. n. 3, Etuliya (middle bed).

Table 4. Mean rate of evolution and life spans of species in *Promimomys* and *Pliopotamys–Ondatra* lineages

Promimomys			*Pliopotamys–Ondatra*			
Species	Life spans of species, ka	Rate of evolution (number of species per ka)		Species	Life spans of species, ka	Rate of evolution (number of species per ka)
Promimomys sp. n. 1 (Vinogradovka)	1530	0.00065	I	*Pliopotamys minor*	1142	0.00088
P. sp. n. 2 (Mugureny)	1050	0.00095	II	*P. meadensis*	906	0.00110
P. sp. n. (theoretically predicted)	750	0.00133	III	*Ondatra idahoensis*	631	0.00158
P. insuliferus	500	0.00200	IV	*Ondatra* of the zone IV (Martin, 1984)	454	0.00220
P. moldavicus	300	0.00333	V	*O. annectens*	322	0.00311
P. sp. n. 3 (Etuliya, Lucheshty)	120	0.00833	VI	VI theoretically predicted	222	0.00450
			VII	*O. nebrascensis*	149	0.00671
			VIII	theoretically predicted	92	0.01087
			IX	"	50.5	0.01980
			X	*O. zibethicus*	22.4	0.04464

noted that the given figures of rates are averaged, because evolution in every lineage and in each interval of a curve is accelerated. Each lineage shows a minimum rate of taxa origination in first forms and a maximum one in terminal forms (Table 4).

The position of different forms on the inverse hyperbolic cotangent curve allows the comparison of relative stages of evolutionary advancement of forms from different phyletic lineages (Fig. 1). For example, the position of a species from one phylum, situated in a more gentle part of the curve, as compared with the species of another phylum testifies to the lesser evolutionary advancement of the former than the latter species in its lineage. Rate of evolution of the former species can be higher than that of the latter one.

CONCLUSION

It has been established that the evolution of mammals in studied phyletic lineages follows an uniform law that is described by the inverse hyperbolic cotangent formula. This law can be used for solving a number of theoretical and practical problems, such as: bridging the gap between paleospecies and neospecies concepts, determining morphological limits of species and their life spans, determining evolutionary rates, and extensive application of the biometric method for the precise dating of fossil remains.

Analyzed data implies: (1) All phyletic lineages display an accelerated change of morphological evolutionary parameters through the course of time; (2) Species life spans decrease up the time scale; (3) If the evolutionary rate is the number of taxa originating during a time unit, then evolutionary rate increases up the time scale.

The present study was supported by grants from the International Science Foundation and the Academy of Natural Sciences.

Reviewers A.K. Agadzhanyan and A.G. Ponomarenko

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Dynamics of the Global Diversity of Proterozoic Stromatolites. Article I: Northern Eurasia, China, and India

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Received December 1, 1993

Abstract – Dynamics of the global diversity of Proterozoic stromatolites are analyzed at species and generic levels on the basis of separate considerations of the data on each of the six largest superregions (the data on three of them are discussed herein), with subsequent generalizing investigations on superregional samples of stromatolite taxa. This approach allowed us to minimize statistical noise attributable to the inadequate understanding of volumes of taxa of the same rank by specialists from different schools, as well as to reveal the influence of different-scaled geological processes on stromatolite diversity changes. The abundance of taxa was analyzed for 11 time intervals of comparable duration. Four of them were delimitated in the Early Proterozoic, and seven were recognized in the Late Proterozoic. The most representative samples of stromatolite taxa, derived from northern Eurasia and China, show similar dynamics of change during the Proterozoic, as illustrated by their very similar histograms of the diversity of species and genera. Both pairs of histograms have a three-modal character and demonstrate the same maxima and minima positions as well as a general sharp diversity decrease during the second half of the Late Riphean (0.85 - 0.65 Ga ago), which continued in a smoothed form during the Vendian. The sample of Proterozoic stromatolites from India is less representative, because it only characterizes the well studied areas. Nevertheless, the relevant diversity histogram, though evidently peculiar, bears some similarity with those for northern Eurasia and China. The geological data implies a relation between the diversity variations of pre-Vendian stromatolites and the significant sea transgressions (maxima of diversity) and epochs of folding and regressions (minima of diversity). It also reveals a specific Vendian decline of stromatolites during a wide postglacial transgression in northern Eurasia and southern China. The similarity in the dynamics of diversity of Proterozoic stromatolites in these superregions suggests that these dynamics were controlled by the events and processes common for at least the majority of Eurasia.

INTRODUCTION

Stromatolites are an important source of biogeological information on the long interval of the Earth's Precambrian history preceding the Vendian diversification of Metazoa. Stromatolites occurred sporadically in the Archean, particularly in the Early Archean; in the Proterozoic, they reached their maximum abundance, morphological diversity, and ecological distribution for the whole Earth's history. Diagnostic features of stromatolite buildups as biosedimentary structures reflect a complicated interaction between communities of microorganisms, predominantly Cyanobacteria, and the environment. The mechanism of this reflection remains yet unclear. However, there are strong grounds for believing that the buildup microstructure, which is the main or single criterion for recognizing formal species (forms) of stromatolites, and, to some extent, the laminae shape are closely related to the taxonomic composition of stromatolite-forming mats and often allow inferences about the shape, size, and arrangement of cells in the mats (Serebryakov, 1975; Gebelein, 1974, 1976; Semikhatov *et al.*, 1979; Walter, 1983; Awramik, 1991, 1992).

Understanding the biogeological informativeness of stromatolites, many researchers, beginning with Maslov (1959), attempted to evaluate variations in their abundance and diversity with time. This was done first on qualitative (for bibliography see Semikhatov and Raaben, 1993), and then on quantitative bases (Awramik, 1971, 1991, 1992; Walter and Heys, 1985; Zhu, 1982; Coa, 1988; Semikhatov and Raaben, 1993). Analysis of diversity dynamics was commonly based on the quantity of taxa in stratigraphic units that had lasted some hundreds of millions years. The integrated data on long intervals reduce the influence of the facts that: the main stromatolite-bearing strata occur discretely in the sections; their intercontinental, or even interregional correlations are often far from being clear; and the duration of their formations is commonly unknown. Unlike most authors, Cao (1988) tried to derive the diversity dynamics of Proterozoic stromatolites from the quantitative variations of species in relatively short time intervals (10 Ma). Although this approach is typical when studying Phanerozoic fossils, it is unlikely that it is correct for the Proterozoic (Walter and Heys, 1985) because of the incomparable degree

of stratigraphic resolution and accuracy of telecorrelations in the Precambrian and Phanerozoic, as well as the great time ranges of many stromatolite taxa.

The informative value of any model for variations in the diversity of fossils depends largely on the time resolution of the stratigraphic framework and the accuracy of telecorrelations between fossiliferous strata. The models of Awramik (1971, 1992) and Walter and Heys (1985) used, in fact, the same stratigraphic framework (Fig. 1). To analyze the Riphean and Vendian stromatolites of northern Eurasia, we addressed the problem in another way (Semikhatov and Raaben, 1993). By doubling the time resolution of the Riphean portion of the scale, we recognized some new peculiarities in the diversity changes and came to the conclusion that existing notions on the time and causes of the Late Proterozoic stromatolite decline need to be revised.

Another factor determining the informative value of the above models is the representativeness of the samples used. Figure 1 shows that the general trend of the diversity changes in the Proterozoic stromatolites (an increase from the beginning of the Proterozoic toward its second half, and a sharp decline in the Vendian and Cambrian) is clearly indicated by samples of different volumes (70, 485, and 663 taxa). However, the mono- or bimodal character of the trend depends on the size of the samples used (Figs. 1B and 1C). This, therefore, presents questions concerning the extent to which corresponding histograms show the real diversity changes rather than the state of our knowledge (Awramik, 1992). In this connection, we should note the following.

First, the diversity dynamics can be established correctly by analyzing the taxa quantities in time intervals of comparable length only. In the above case (Figs. 1B and 1C), the taxa frequency in the Late Aphebian was calculated for an interval twice as long as the intervals of the Riphean. This disproportion could not avoid having an effect on the quantities of the taxa found.

Second, two-thirds of the new taxa estimated in Awramik's histogram (Awramik, 1992; Fig. 1C) are confined to two short portions of the Proterozoic section in China and were established by specialists (Liang *et al.*, 1984; Zhu *et al.*, 1987) who treat the volumes of stromatolite taxa much more narrowly than researchers outside of China.

The aforesaid reflects the general unsatisfactory state of the classification and taxonomy of stromatolites. The presence of indiscerned synonyms and invalid taxa is a part of the problem. Another, and probably more important part is that taxa of the same rank, identified by different specialists, are often inadequate in their sets of diagnostic features and their volumes, and they sometimes overlap each other (for review and bibliography see Krylov, 1975; Semikhatov and Komar, 1989). Thus, some researchers (mainly Russian) are in favor of the strict ranking of stromatolite diagnostic features. These researchers usually distinguish formal genera according to a combination of morphological peculiarities of buildups, and formal species,

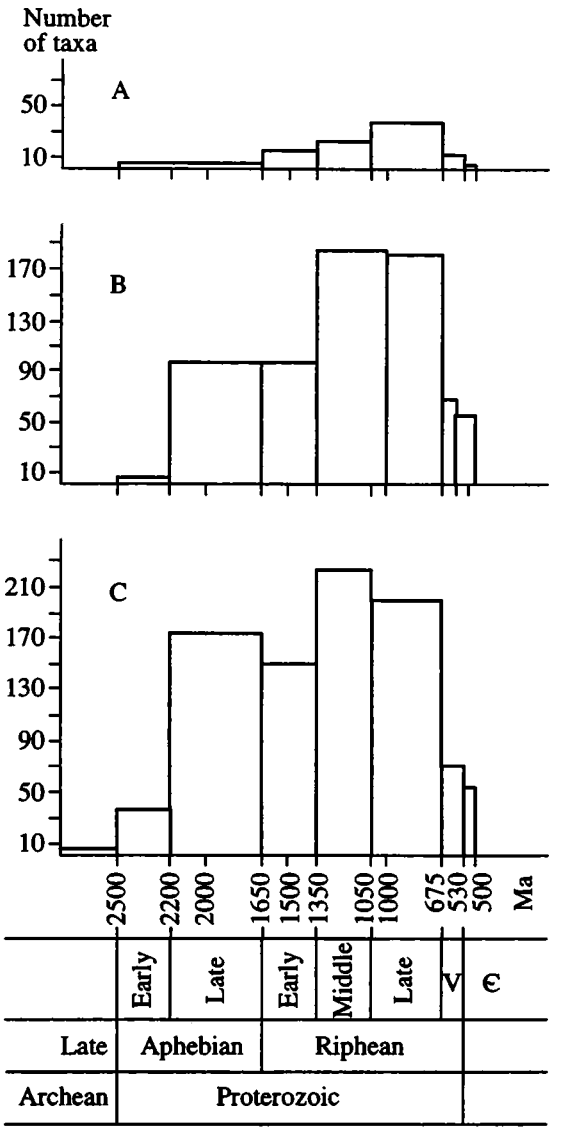

Fig. 1. Dynamics of global stromatolite diversity in the Proterozoic and Early Cambrian according to Awramik, Walter, and Heys.

A – according to Awramik (1971); B – according to Walter and Heys (1985); C – according to Awramik (1992).

exclusively or predominantly by the microstructure. Other researchers (e.g., Bertrand-Sarfati, 1972; Walter, 1972; Preiss, 1972, 1973; Grey, 1984) substantiate their taxa by a combination of any recognizable features, both macro- and microscopic ones, allowing for a difference in the taxonomic value of a feature of morphologically different stromatolites. Therefore, formal species (forms) of some formal genera (groups) were established by the microstructure and others, by the laminae shape or the morphological peculiarities of buildups; some genera were distinguished mainly according to their microstructure, whereas most genera were established regardless of the microstructure, and so on. Lastly, the third group of researchers (particularly the Chinese) supports a narrow understanding of

stromatolite species, distinguishing them by microstructural peculiarities of minor importance and/or buildup morphological features, and substantiating genera by the combination of the latter. As a result, the statistical estimations summarizing all stromatolite data are burdened with casual errors that are determined by the relationships between the quantities of taxa established in the given time interval by specialists of different schools.

A strict way to solve this problem could be a monographic revision of the type materials of numerous close stromatolite taxa; however, this is not feasible. Nevertheless, we discerned approximately one dozen synonymic forms by comparing their type species with the available holotypes of other taxa first identified earlier.

Taking into consideration the aforementioned, we conducted a separate statistical analysis of the taxa established by specialists of different schools and derived from each of the six largest superregions (northern Eurasia, China, India, Africa, Australia, and North America). For the purpose of analyzing the diversity dynamics for any fossil group, the meaningful parameter is not the absolute quantity of taxa in a time interval, but the variations in taxa diversity in two neighboring intervals, i.e., the relative values. A separate study of the data provided by adherents of different approaches to recognizing stromatolite taxa allows us to eliminate in each superregional sample the influence of the above-mentioned casual error, which is related to the inadequate volumes of taxa of the same rank. Thus, each sample appears to be burdened only with a systematic error, which results from the understanding of the taxa volumes by representatives of a single school and cannot affect estimations of the diversity changes in time.

The separate data analysis of each of the superregions has another advantage: it more accurately reveals the relations between the changes of stromatolite diversity with time and the specific features of the geological history of the territory. This was first illustrated by the example of northern Eurasia (Semikhatov and Raaben, 1993).

The stratigraphic framework of the present analysis was the refined Precambrian stratigraphic scale for northern Eurasia within the limits of the former USSR, adopted in 1990 (Semikhatov *et al.*, 1991). The Lower Proterozoic (Karelian) of this scale is almost equal in time range to the Palaeoproterozoic of the Precambrian scale that was suggested by the International Subcommission on Precambrian Stratigraphy (Plumb, 1991) (2500 ± 50 - 1650 ± 50 and 2500 - 1600 Ma, respectively). For our purpose, we divided this interval into four units with inner boundaries of 2300, 2000, and 1800 Ma. These units actually coincide with the Palaeoproterozoic subdivisions of the International Subcommission scale (Siderian, Rhyacian, Orosirian, and Stetherian). The differences are that the termination of the fourth unit is dated as 1650 ± 50 instead of 1600 Ma, and the age of the boundary between the second and third units is 2000, not 2050 Ma. This boundary position was determined by biospheric evolutionary events, whereas the boundary of the International Subcommission scale was based on the age of the endogenic processes.

In the Riphean–Vendian part of the Proterozoic (in the Meso- and Neoproterozoic), we used a scale that envisaged a recognition of three Riphean erathems (the Lower, Middle, and Upper Riphean; or the Bursyanian, Yurmatinian, and Karatavian) and the Vendian System, as well as the informal twofold subdivision of each of the Riphean erathems. The isotopic ages of the lower boundaries of the Lower, Middle, and Upper Riphean and Vendian are known to be 1650 ± 50, 1350 ± 20, 1000 ± 50, and 650 ± 20 Ma, respectively (Semikhatov *et al.*, 1991). The boundaries between the subordinate Riphean units are dated as ca. 850, 1200, and 1450 Ma (Semikhatov and Raaben, 1993).

The interregional and intercontinental correlation of these units was based both on the available isotopic-geochronological data, which varies in methodical substantiation and reliability, and on the composition of stromatolite assemblages occurring in the successions under comparison. In some cases, however, there were considerable difficulties in the correlation, which were caused by a lack of isotopic dates and an endemism of the stromatolite assemblages. Therefore, there is sometimes no certainty that we traced strictly isochronous units. However, according to the available data, a possible or probable time-transgression of the boundaries of these units is small in comparison with the duration of these units and can be neglected here.

When analyzing the data on the superregions, we counted the taxa, the identification of which met every necessary requirement. The varieties described in the literature were regarded as independent species. The taxa defined in the open nomenclature, and especially those identified by field observation, were not counted. The inner structure of the representative samples was analyzed with regard to the relationships between high-rank (supergeneric) morphological groupings of buildups. In contrast to the previous study (Semikhatov and Raaben, 1993), columnar branching buildups were taken as a single unit.

For purely technical reasons, the results obtained are discussed in two articles. The first article defines the problem, summarizes earlier data on the diversity of northern Eurasian stromatolites, and presents original material on the diversity dynamics for China and India stromatolites and the correlation of these dynamics with the superregional geological events. The second article discusses the analogous material on stromatolites of Africa, Australia, and North America and compares the superregional samples. Changes in the global diversity of Proterozoic stromatolites are analyzed and related with the major abiotic and biotic events. The time and probable causes of the stromatolite decline in the Late Proterozoic are discussed as well.

DYNAMICS OF DIVERSITY OF PROTEROZOIC STROMATOLITES IN THE SUPERREGIONS

Northern Eurasia

In northern Eurasia, Precambrian stromatolites have a nonuniform distribution in time and space. They are unknown from the Archean. Early Proterozoic stromatolites occur in some areas of the eastern Baltic shield and in three localities of the Asiatic part of the superregion, where they constitute relatively thin carbonate bodies that are strongly subordinated to volcanogenic-terrigenous strata, being small fragments of the general succession. In the Upper Proterozoic, the stromatolites compose thick and extensive members, which occur more or less uniformly in the Riphean sections and dominate them locally. The Vendian stromatolites show infrequent occurrences. By the lateral distribution and the degree of study of their taxonomic composition, the Late Proterozoic stromatolites are first among contemporaneous biolites from the other superregions.

The diversity dynamics of the Proterozoic stromatolites of northern Eurasia can be seen in Fig. 2. Here, as well as in the following figures, the histograms show the absolute quantities of formal species and genera in the units of the accepted time scale, and the circle diagrams demonstrate the inner structure of each age sample – the percentage of taxa belonging to different supergeneric morphological categories. The difficulties in relating the Riphean and Vendian stromatolite-bearing strata of northern Eurasia to the accepted scale, which were significant for our analysis, were discussed earlier (Semikhatov and Raaben, 1993). In the Lower

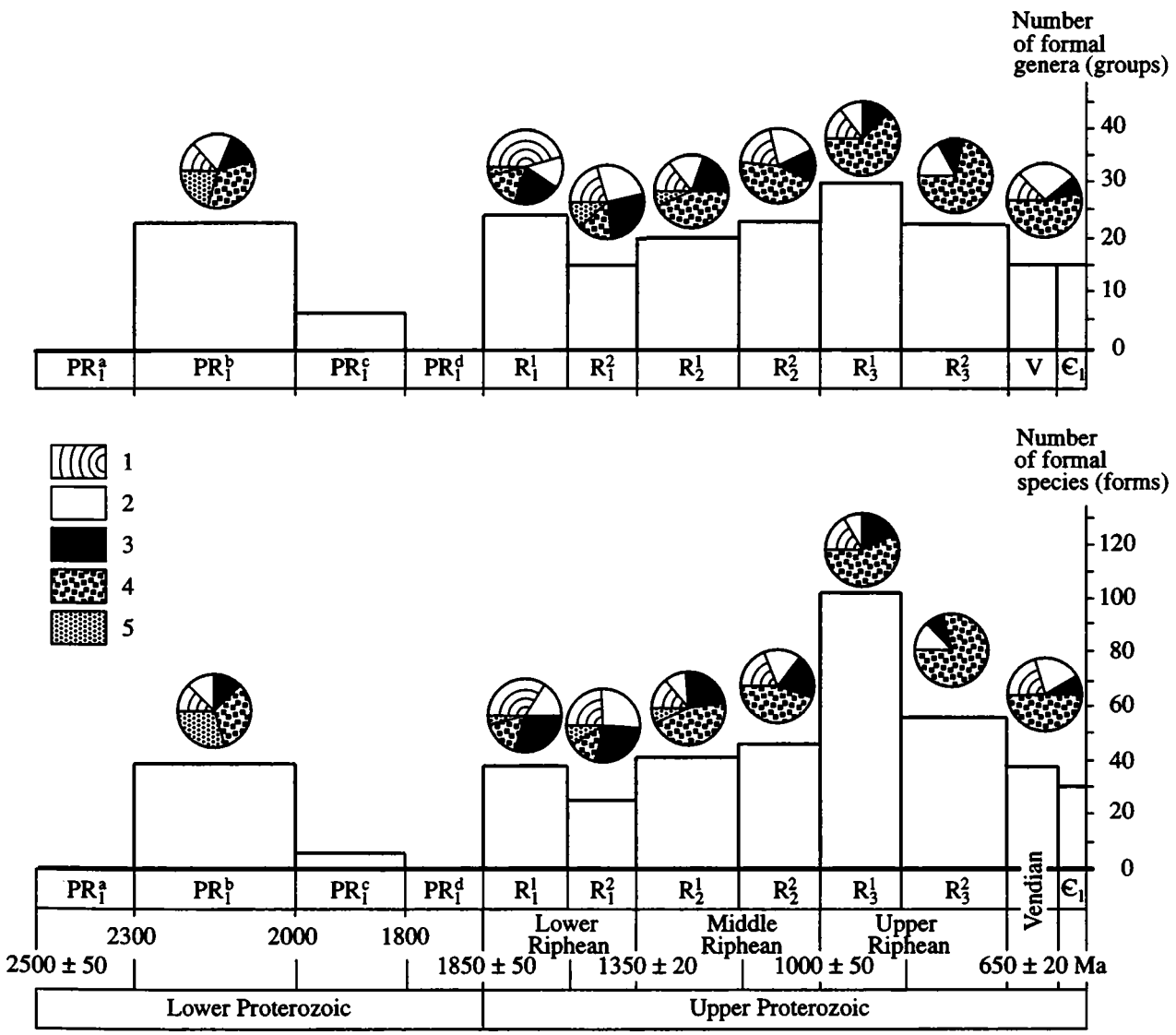

Fig. 2. Dynamics of the taxonomic diversity of the Proterozoic and Early Cambrian stromatolites in northern Eurasia at the levels of formal species (forms) and formal genera (groups); after Semikhatov and Raaben (1993), with amendments.

1 - 5 – percentage of the stromatolite species and genera belonging to supergeneric categories: 1 – stratiform morphs, 2 – nodular morphs, 3 – columnar unbranching morphs (conophytonids), 4 – columnar branching morphs, and 5 – microstromatites.

Proterozoic part of the section, the assignment of the fossiliferous beds to the scale was reliably substantiated by the U-Pb zircon datings of volcanics alternating with the mentioned beds and relevant intrusions.

The plot of the changes in the diversity of the Proterozoic stromatolites of northern Eurasia, presented in Fig. 2, is somewhat different from the previously published one (Semikhatov and Raaben, 1993, Fig. 1). Differences in the Lower Proterozoic part of the section resulted from using partly different time-units, and those in the Upper Proterozoic and Cambrian sections are related to the widened database. Figure 2 includes newly published data (Dol'nik *et al.*, 1990; Kyrvel, 1991) and some earlier materials, which unfortunately escaped our attention (Golovanov, 1972; Dol'nik, 1978 a, b; Mal'tsev *et al.*, 1982). Most of the data on the distribution of stromatolite taxa in the Proterozoic and Lower Cambrian of Northern Eurasia was obtained, as earlier, from summarizing reviews and publications containing paleontological descriptions: Korolyuk, 1960; Sidorov, 1960; Semikhatov, 1962; Krylov, 1963, 1967, 1975; Raaben, 1964, 1969, 1991; Komar *et al.*, 1965; Komar, 1966, 1990; Nuzhnov, 1967; Golovanov, 1967, 1970; Rozanov *et al.*, 1969; Semikhatov *et al.*, 1970; Korolyuk and Sidorov, 1971; Raaben and Zabrodin, 1972; Khomentovskii *et al.*, 1972; Dol'nik and Vorontsova, 1974, 1984; Dol'nik and Nikol'skii, 1974; Shapovalova, 1974; Shenfil', 1975, 1991; Shpunt *et al.*, 1976, 1979, 1982; Vlasov, 1977; Krylov and Perttunen, 1978; Lyubtsov, 1979; Bertrand-Sarfati and Siedlecka, 1980; Dol'nik, 1982; Stratotip ..., 1982; Semikhatov and Serebryakov, 1983; Makarikhin and Kononova, 1983; Medvedev, 1991; Raaben, 1991.

The sample of Proterozoic and Early Cambrian stromatolites of northern Eurasia consists of 335 species of 86 formal genera. In the Lower Proterozoic, there occur 44 species and 24 genera; in the Upper Proterozoic and Lower Cambrian, 293 species and 72 genera occur, including 2 species and 10 genera passing from the Lower Proterozoic. The sample became more representative than the previously published one at the generic level, at the expense of the Cambrian taxa and, at the specific level, at the expense of the forms from the Upper Riphean deposits of the Transbaicalian region, the extreme northeastern area of Asia and Taimyr. It should be noted that the time-constraints of these deposits were based mostly or entirely on the enclosed stromatolites and are sometimes treated ambiguously by different workers.

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The Riphean–Vendian segment of the histogram is divided into three distinct parts: (1) the Early-Middle Riphean part, characterized by a general gradual growth of diversity with a slight drop in the Late Burzyanian, which resulted in the isolation of the Early Burzyanian relative maximum; (2) the Early Karatavian part (1000 - 850 Ma), representing a drastic peak of diversity, especially at the specific level; and (3) the Late Karatavian–Vendian part, marked by a general decline in diversity, which was the most drastic in the Late Karatavian and more smoothed in the Vendian. This decline manifested itself in the noticeably reduced number of newly appearing taxa and the more considerably increased number of disappearing taxa, especially at the transition from Early to Late Karatavian.

In the course of all quantitative changes, the most dynamic components of the stromatolite assemblages appeared to be the taxa of columnar buildups. It was the variations in their quantity that provided the diversity changes at both specific and generic levels. The role of columnar stromatolite taxa was increasing steadily during the Riphean at the expense of layered and nodular ones; it decreased during the Vendian and Cambrian. This was accompanied by the directed redistribution of the relative roles of different supergroup categories of columnar buildups (Fig. 2, circle diagrams; Semikhatov and Raaben, 1993, Figs. 2 and 3).

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Among the abiotic factors responsible for variations in the abundance and diversity of stromatolites in the superregion, the paleogeographic factor – expansion and reduction of paleobasins, variations in character and environments of sedimentation, and other changes – is the most evident. The general Late Proterozoic increase in the abundance and diversity of stromatolites surely reflects the appearance of more extensive stable massives and, correspondingly, shallow-water platform basins during that time than in the Early Proterozoic (Semikhatov, 1974; Borukaev, 1985). It will suffice to remember that the gigantic cratons of northern Eurasia were formed in the post-Svekofennian time, and the first transgressions covered them in the Riphean time.

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Proterozoic part of the section, the assignment of the fossiliferous beds to the scale was reliably substantiated by the U-Pb zircon datings of volcanics alternating with the mentioned beds and relevant intrusions.

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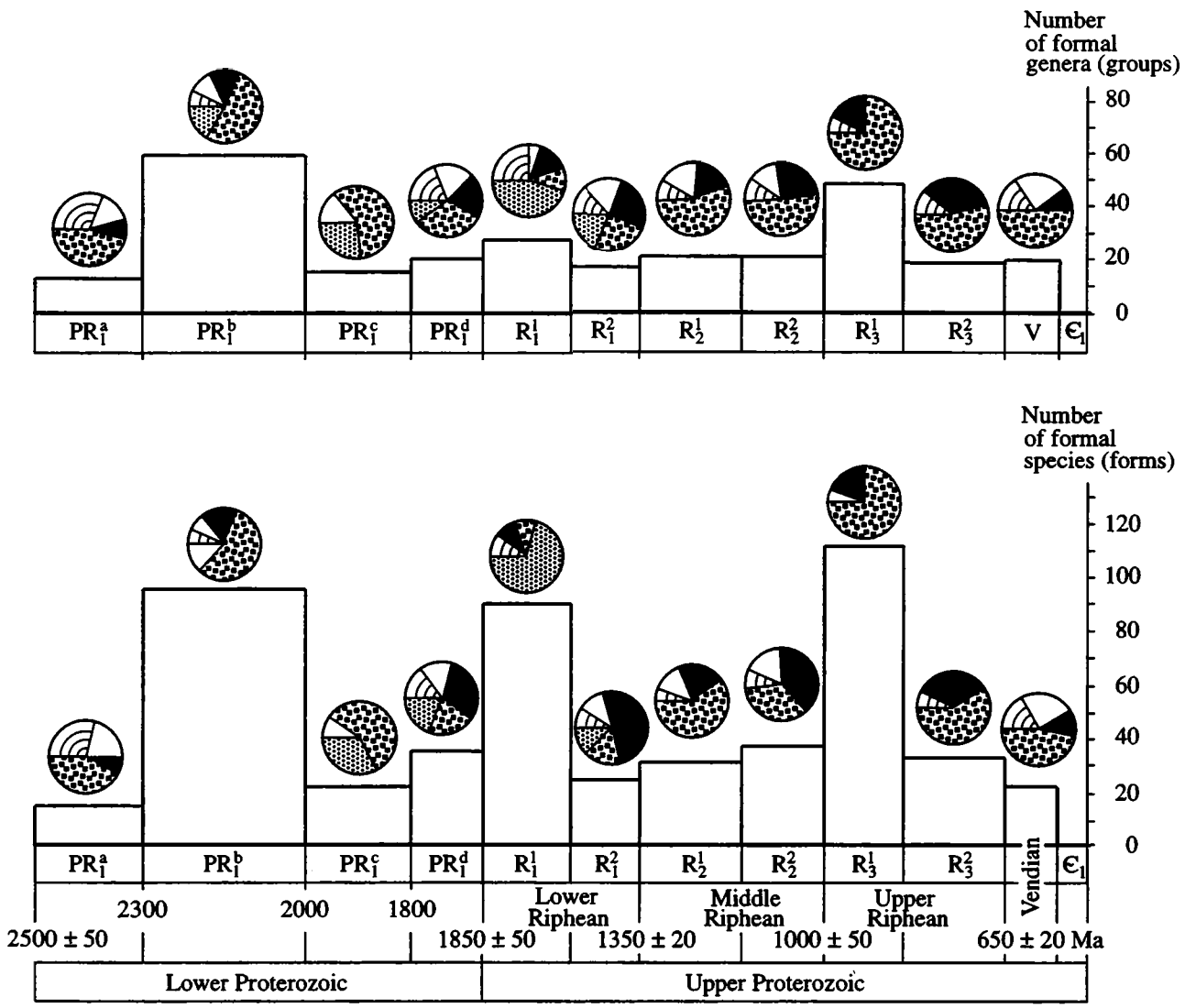

Fig. 3. Dynamics of the taxonomic diversity of the Proterozoic stromatolites in China at the levels of formal species (form) and formal genera (groups). For symbols see Fig. 2.

are known outside the country, and a significant part of the taxa is more endemic in origin. The diversity and high endemism of low rank taxa do not seem to reflect the situation adequately, but they can be attributed to the narrow volume of stromatolite taxa taken by the Chinese specialists. Sixty-two species and 12 genera of the Pseudogymnosolenacea "family," first identified by Liang and others (Liang *et al.*, 1984), can serve as an example. Many of these species have a very limited lateral distribution and, according to the latest view of Prof. Liang (pers. commun., October 1992, the town of Tianjin), should be considered as synonyms.

However, it should be remembered that the occurrence of indiscerned synonyms cannot considerably affect the general pattern of diversity changes if the stromatolites are studied by specialists of a single school. In this case, according to statistical laws, the number of synonyms varies proportionally with the sample size, thus amplifying the fluctuations in passing from one age sample to another, but it cannot change the sign of the fluctuations.

When analyzing the stromatolites of China and their age assignments, we used both the publications in English (Zhu, 1982; Cao, 1983; Liang *et al.*, 1985; Zhang, 1985; Zhu and Chen, 1992) and the available original works (Zhu, 1980; Cao and Liang, 1974; Liang and Tsao, 1976; Cao and Zhao, 1977, 1981, 1985; Geological Guide, 1977; Tsao (Cao), 1979; Cao, 1980, 1983, 1992; Liang, 1980, 1992; Zhang *et al.*, 1980, 1992; Liang *et al.*, 1984, 1987; Lin *et al.*, 1984, 1987; Miao *et al.*, 1984; Zhao X. *et al.*, 1984, 1987; Zhao W. *et al.*, 1985; Xing *et al.*, 1985; Dai and Peng, 1987; Gong, 1987; Li and Leng, 1987; Wang and Gao, 1987; Xu *et al.*, 1987; Zhu and Wu, 1987; Zhu *et al.*, 1987, 1992; Hua and Qiu, 1992; Qiu, 1992; Qiu and Hua, 1992a, b; Xiao, 1992; Zhang and Wang, 1992; and others). Certainly the language barrier was a great difficulty, because there was an English resume only in

some of these publications. Of help were illustrations, plates, and diagrams, which showed the distribution of individual taxa throughout the stratigraphic units. Names of the latter written in hieroglyphs were commonly recognizable. We failed to define the stratigraphical assignments of only 13 forms described in the Chinese publications. These forms were not analyzed, as well as a small set of forms occurring in the correlatives of the Yinmin Formation from the Shennongjia region, the western Hubei Province. The stratigraphic position of these deposits is disputable, and the set of stromatolite taxa consisting only of Riphean genera does not support the pre-Riphean age of the enclosing deposits.

The available literature provided us with information on the distribution of 367 forms representing 141 groups of stromatolites in the Proterozoic of China and North Korea. In the sample we included 48 taxa of species level at the expense of the morphotypes that had been identified as genera only but were the single representatives of those genera in the definite units of the accepted scale. When analyzing the inner structure of the samples, we corrected the volumes of some supergeneric groupings of the Riphean–Vendian buildups identified by the Chinese specialists, because they united stromatolites of sharply different morphology.

The stratigraphic aspect of the analysis of the Chinese materials presented two problems: the intraregional correlation of the stromatolite-bearing deposits and their relation to the accepted scale. The Lower Precambrian strata of China (older than 1.85 - 1.90 Ga) contain stromatolites in a relatively small area. Thus, their intraregional correlation based on direct geological and, partly, paleontological data is not difficult (Zhu *et al.*, 1987; Zhu and Chen, 1992). They are well dated, mostly by the U-Pb zircon method. As for the Upper Precambrian, which is subdivided into the Changcheng, Jixian (Chihsien), Qingbaikou, and Sinian "systems," it is characterized by stromatolites occurring over broad territories (Liang *et al.*, 1985, Fig. 1). The intra-China correlation of these deposits is based on: the isotopic-geochronological determinations yielded by the K-Ar, Rb-Sr, and Pb-Pb dating of the sedimentary geochronometers; the occurence of easily traceable successive stromatolite assemblages; and some geochemical criteria. In most cases, the correlations are not a subject of significant controversy. The exception is the upper horizons of the Upper Precambrian of the eastern North China platform, distributed in the Liaodong Peninsula and the Anhui Province. These horizons were correlated to the accepted scale on the basis of isotopic-geochronological data, taxonomic composition, and succession of their stromatolite assemblages. The difficulties of this correlation are discussed below.

The oldest (2.5 - 2.0 Ga) stromatolite-bearing deposits of China, known as the Hutuo Group and its correlatives, have a limited distribution in the riftogenic structures of the southern Shanxi Province (Wang and Qiao, 1984; Zhu *et al.*, 1987). The first stromatolites appear in the Doucun Subgroup, the lower part of the Hutuo Group, which unconformably overlies the Archean rocks (2520 ± 17 Ma, and older). Volcanics at the top of the subgroup yielded the U-Pb zircon dates of 2366 + 103/-94 Ma (for bibliography see Zhu and Chen, 1992). The subgroup contains a small set of taxa (Fig. 3) (Zhu *et al.*, 1987) dominated by interprovincial genera of layered, nodular, and columnar buildups. Most genera of the latter do not occur beyond the Lower Proterozoic. Species of the columnar stromatolites are predominantly endemic, but the species composition of other types of buildups needs more exact definition.

In the interval of 2.3 - 2.0 Ga, the number of stromatolite genera and species is several times larger than in the preceding interval. This sharp outburst of diversity is at the expense of the taxa from the middle horizons of the Hutuo Group, the Dongye Subgroup, and also from its correlatives – the Gantaohe and Zhongtiao Groups (Zhu *et al.*, 1987; Zhu and Chen, 1992). Scarce taxa from the Dashiqiao Formation of the Liaohe Group in the eastern Liaoning Province (Cao and Zhao, 1977, 1981) belong probably to the same time interval. However, they are of minor importance for the general pattern of diversity changes of the Early Proterozoic stromatolites in China. This group is commonly correlated with the Hutuo Group (Wang and Qiao, 1984; and others) and sometimes with the lower part of the Changcheng "System" (Zhong *et al.*, 1984). We are in favor of the first correlation, which is supported by the stromatolite content.

The stromatolites from the Dongye Subgroup and Gantaoche Group constitute some successive local assemblages, the diversity of which grows from the basal horizons upward the section but sharply decreases in the upper formations. The granites that intrude the Gantaoche Group yielded the U-Pb zircon age of 2056 ± 12 Ma. The maximum age of the Dongye Group is estimated as 2300 Ma, according to the above U-Pb dating of the underlying volcanics of the Doucun Subgroup (Zhu *et al.*, 1987). Thus, the discussed stromatolites reached a diversity peak 2.2 - 2.1 Ga ago.

The increase of stromatolite diversity in China 2.3 - 2.0 Ga ago was provided by columnar branching buildups and the appearance of microstromatites. Among the first, there are many taxa that are referred by Chinese specialists to some supergroups ("families") widespread in the Riphean (Kussiellidae, Tunpussidae, Gymnosolenidae) (Zhu *et al.*, 1987). But according to the published reconstructions, at least a portion of these taxa should be referred to the supergroup of irregularly branching morphs (intricatids) (Raaben, 1981, 1986), spread in the Lower Proterozoic of Karelia and Canada.

In the genera set from this time interval, an appreciable role belongs to the interprovincial taxa. In addition to those having wide time ranges, the set includes the genera characteristic of the Lower Proterozoic (e.g., *Djulmekella*, *Discorsia*, *Gruneria*, and *Pilbaria*) and some genera confined to the particular Riphean horizons in northern Eurasia. Identifications of the latter in the Hutuo Group and the overlying Lower Proterozoic

horizons are often doubtful. The species composition of the stromatolites from these deposits is dominated by narrow endemics. In addition, there are *Djulmekella djulmekensis* Mak., *Eocapsiphora paradisia* Cloud et Semikh., *Pilbaria perplexa* Walt., and others recorded from the Lower Proterozoic outside of China.

There are no stromatolites in the upper subgroup of the Hutuo Group and its correlatives composed of terrigenous deposits. However, they are widespread in the overlying deposits of the Changcheng, Jixian, and Qingbaikou "systems," which fill some avlacogens, compose basal horizons of the cover of the North China and Tarim platforms, and occur as constiuents of the Yangtze platform basement (Wang and Qiao, 1984; Goodwin, 1991).[1]

The oldest of those stromatolites are known from the iron-bearing member at the base of the Chuanlinggou Formation (the second one from the base of the Changcheng "System" in the Yanshan Range). This member contains an assemblage of local species of *Eucapsiphora, Gruneria, Nucleella, Omachtenia,* and other genera, including the Chinese endemics. Representatives of the two first genera and of *Nordia* and *Cryptozoon* occur in the upper carbonate part of this formation (Zhu and Chen, 1992). A similar assemblage was recorded from the Dayanping Formation in the Shennongjia region, the Yangtze platform (Li and Leng, 1987; Zhu and Wu, 1987). The association of stromatolites from these units of two isolated regions differs from the preceding one in the reduced specific and generic composition and the absence of stratiform and columnar unbranching forms.

Although the stratigraphic position of these stromatolites is clear (in the lower part of the Changcheng "System"), their isotopic age is estimated ambiguously because of the different datings of the lower boundary of this "system" in the Chinese literature (from 2000 to 1800 Ma). The maximum estimations were based on the Pb-Pb and U-Pb isochronous dating of whole-rock samples of black shales from the Chuanlinggou Formation (1922 ± 91 and 1910 ± 70 Ma) (Chen *et al.*, 1981; Zhang and Zhang, 1985). Considering the recent data, the geological meaning of these datings is very doubtful. The following facts reinforce that doubt. First, a fine illite fraction from the Chuanlinggou and the underlying Changzhougou formations of the Jixian "System" yielded much younger Pb-Pb isochronous and K-Ar dates – 1785 ± 19 and 1848 ± 39 Ma, respectively. Second, the U-Pb concordant and Pb-Pb isochronous ages of apatite from the metamorphic rocks of the Taihang Mountains (the Dongzhao Group), unconformably overlain by the Changzhougou Formation, are 1810 and 1802 ± 68 Ma (for bibliography see Zhu and Chen, 1992).

According to these data, the entire Changcheng "System" may be younger than 1800 Ma, because the U-Pb and Pb-Pb apatite datings of the metarocks fix the time of completion of the metamorphic processes (Ovchinnikova *et al.*, 1993). However, the Chuanlinggou Formation is cut by lamprophyres yielding a K-Ar biotite date of 1800 Ma and a greater date for the whole rock samples. These and the mentioned Pb-Pb illite datings probably were regarded by the Chinese authors' as the basis for estimating the age of the stromatolites from the Chuanlinggou Formation as 1850 - 1800 Ma (Zhu *et al.*, 1987; Zhu and Chen, 1992). We accept this estimation, although the above data suggests that the stromatolites may be younger than 1800 Ma. However great the isotopic age of the Chuanlinggou and Dayanping stromatolites may be, the interval of 2.0 - 1.8 Ma is the poorest one of the Proterozoic of China in stromatolite composition (Fig. 3).

The stromatolites from the overlying Precambrian horizons constitute several successive assemblages (assemblages I - VIII), persistent in their composition throughout vast areas and widely used for correlation of the enclosing deposits (Liang *et al.*, 1985; Xing *et al.*, 1985; and others).

In the terminal interval of the Lower Proterozoic scale (1.8 - 1.65 Ga) there occur: (a) the stromatolite assemblage I, which is reported in the Yanshan type section from the Tuanshanzi Formation and the lower part of the Dahongyu Formation of the Changcheng "System" and (b) a few forms from the upper horizons of the Dahongyu Formation (Zhu and Chen, 1992), which were included earlier in assemblage II. Carbonate rocks of the first formation have the Pb-Pb isochronous age of 1776 Ma (Chen *et al.*, 1981), and volcanics of the second formation have the U-Pb zircon age of 1621 ± 6 Ma (Lu and Li, 1991). The occurrence of some typical taxa allows recognition of the impoverished assemblage I in correlatives of the Changcheng "System" in northwestern China (Zhang *et al.*, 1980; Zhao *et al.*, 1984, 1987; Liang *et al.*, 1985) and possibly in the Liaohe Formation of the Lower Kunyang Group in the Central Yunnan, which has a Pb-Pb isochronous age of 1790 - 1760 Ma (Huang *et al.*, 1987).

The whole composition of stromatolites from the interval of 1.8 - 1.65 Ga is more diverse than the preceding one. This corresponds to its wider lateral distribution. All types of buildups are represented in this association. The generic and specific compositions are dominated by the Chinese endemics and taxa of great time range. In addition, there are abundant interregional *Gruneria* (Liang *et al.*, 1985, 1987) represented by several forms, including *G. biwabika* Cloud et Semikh., which is known from the Lower Proterozoic of North America.

According to the isotopic-geochronological and paleontological evidence, the stromatolites from the

[1] There are some different views on the stratigraphic scope of the Changcheng and Jixian "systems" in their type section in the Jixian region (for details see Brezhnev and Raaben, 1992; Zhu and Chen, 1992) Herein, we follow the Chinese geologists who draw the boundary between these "systems" in the stratotype at the contact of the Dahongyu and Gaoyuzhuang formations, because the upper formation shows an unconformity at its base and a close similarity with the overlying strata of the Jixian "System" in lithology, environments of deposition, and paleontological content (Zhu and Chen, 1992).

lower part of the Jixian "System" of the accepted scope can be assigned to the Lower Riphean. The Chinese authors include these stromatolites in assemblages III (Mopanyu) and IV (Shanpoling) and in the part of assemblage II that is enclosed in the lower horizons of the Gaoyuzhuang Formation. This part has a limited distribution and consists of few taxa. In addition to the predominant narrow endemics, it contains the interprovincial *Conophyton* (including *C. garganicus* Kor.), *Jacutophyton*, and *Kussiella* genera.

In the type sections, the Mopanyu assemblage occurs near the top of the Gaoyuzhuang Formation and in the lower horizons of the two overlying Yangzhuang and Wumishan formations in the Jixian region. Outside the region, the assemblage is distinguished in the correlatives of these formations in the vast territory of the North China platform, from the Liaoning Province in the east to the bordering areas of the Gansu and Shanxi provinces in the west, as well as in the Central Yunnan, the Yangtze platform. The assemblage, consisting of a variety of mainly endemic taxa, is dominated by microstromatites. Among the accompanying types, stratiform buildups are of prime importance. A number of elements are in common with the Early Burzyanian (Satka) stromatolite assemblage of the Urals. These are *Kussiella* forms, abundant in separate beds of the Wumishan Formation and the stratiform morphs with pectinate-conical lamina, which are included in the Urals in the Thyssagetacea supergroup (Vlasov, 1977; Stratotip ..., 1982) and are known as *Sinoconophyton, Stratoconophyton*, and *Conophyton cancellosum* Liang in China.

The Shanpoling assemblage, confined in the type section to the middle part of the thick (over 3.3 km) Wumishan Formation, has a much wider distribution than the Mopanyu assemblage. In the North China and Tarim platforms, it is spread from the Liaoning Province to the Beishan region. In the Yangtze platform, it is known from the Central Yunnan and the Shennongjia region of the Hubei Province (Liang *et al.*, 1985; Xing *et al.*, 1985; Dai and Peng, 1987). The assemblage contains predominantly large *Conophyton* (*C. garganicus* Kor., *C. lituus* Masl., and *C. shanpolingensis* Liang, similar to *C. cylindricus* Masl.) and nodular buildups with subordinated *Colonnella* (including the Siberian *Colonnella discreta* Kom.), rare microstromatites, and stratiform morphs. In the leading taxa, the assemblage is somewhat similar to the Late Burzyanian stromatolite assemblage known in the Urals and some other regions (Raaben, 1985). As for assemblage V (Tieling) from the abovelying deposits, it is unambiguously considered as the Middle Riphean in age (see below).

The isotopic age estimation of the deposits containing assemblages III and IV was obtained from a few Pb-Pb isochronous datings of carbonate rocks (Zhang and Li, 1982; Chen *et al.*, 1981; Huang *et al.*, 1987). The beds underlying the deposits with assemblage III were dated as 1550 Ma in the Jixian region (the middle part of the Gaoyuzhuang Formation) and as 1600 - 1520 Ma in the Central Yunnan (the Etouchang Formation). The beds enclosing assemblage IV have an age of 1410 Ma (the Luzhiang Formation of the Central Yunnan), and the basal horizons of the overlying Tuangcaoling Formation have an age of 1330 Ma. A model Pb-Pb age of galenite from the Gaoyuzhuang Formation is 1434 ± 50 Ma. It should be remembered that the deposits unconformably underlying this formation have the U-Pb zircon age of 1621 ± 6 Ma (Lu and Li, 1991).

The taxonomic composition of the stromatolites from the lower part of the Jixian "System" and the isotopic age of the enclosing rocks indicate their assignment to the Early Riphean (Burzyanian) and do not support the Middle Riphean age of assemblages III and IV (Liang *et al.*, 1985). The mentioned elements of assemblage II and the entire assemblage III can be assigned to the Early Burzyanian and assemblage IV, to the Late Burzyanian.

If so, the Early Burzyanian was marked in China by a noticeable increase in stromatolite diversity at generic and, especially, specific levels, and the Late Burzyanian, by a considerable decrease (Fig. 3). The increase was provided mainly by the microstromatite taxa (members of the Pseudogymnosolenaceae "family"), the volume of which is interpreted by the Chinese authors very narrowly, as indicated above. Therefore, the amplitude of the Early Burzyanian relative diversity peak is somewhat exaggerated. However, this peak cannot be regarded as an artifact: the pre- and post-Early Burzyanian microstromatites of China are one order of magnitude less diverse than those of Early Burzyanian age, and the stratiform forms also contributed to this peak and the subsequent diversity decrease.

Stromatolite assemblage V (Tieling), characteristic of the upper horizons of the Jixian "System," is referred to the Middle Riphean. This assemblage is widespread in the central and western parts of the North China craton and in the Tarim craton: near the top of the Wumishan Formation and in the Tieling Formation of the Jixian region (the type region); in the coeval deposits of the bordering areas of the Shanxi and Henan Provinces and the Helanshan Mountains; and, further to the west, in the Beishan and Kuruktag regions and the Chinese Tien Shan. The assemblage also occurs in the Shennongjia region of the Yangtze platform (Geological Guide ..., 1977; Liang *et al.*, 1985; Zhao *et al.*, 1987; Li and Leng, 1987). It is missing in the eastern areas of distribution of the Riphean deposits in China (the Liaodong Peninsula, the Anhui and Shandong provinces).

The Teiling assemblage differs greatly from the preceding ones in composition, containing no microstromatites and a small amount of stratiform and nodular buildups; columnar buildups are wide spread and dominated by the branching forms. The leading components of the assemblage are *Baicalia, Chihsienella, Tielingella*, and *Tungussia*, commonly accompanied by *Anabaria, Conophyton, Colonnella*, and closely related *Paraconophyton*, and others. Noteworthy is the presence of some species of these genera (*Baicalia baicalica* Masl., *B. unca* Semikh., *Colonnella cormosa* Kom., *Tungussia nodosa* Semikh., and others) typical of the

Middle Riphean of the Urals and Siberia. Some taxa confined to the post-Middle Riphean in northern Eurasia are also reported in the Teiling assemblage.

Some genera (*Conophyton, Baicalia, Colonnella*, and *Tungussia*) of assemblage V occur throughout its entire time range, but others are restricted to either the lower or upper part of this range, thus dividing the assemblage into two successive sets of taxa. *Anabaria, Chihsienella, Scopulimorpha, Tielingella*, and *Colonnella cormosa* Kom. are spread in its upper part only – in the Tieling Formation of the Yanshan, the upper horizons of the Pingtoushan Group of the Beishan Range, the Bayanxibie Group of Inner Mongolia, the Shiqaohe Formation of the Shennongjia region, and the Dalongkou Formation of the Central Yunnan (Zhang *et al.*, 1980; Liang *et al.*, 1985; Zhao *et al.*, 1987; Li and Leng, 1987; Dai and Peng, 1987). The composition of stromatolites restricted to the lower part of assemblage V varies laterally. Thus, *Wumishanella, Paraconophyton*, and *Pseudochihsionella* dominate in the Yanshan, and *Jacutophyton* and *Cryptozoon* prevail in the Beishan and neighboring regions (Geological Guide ..., 1977; Tsao, 1979; Miso *et al.*, 1984; Zhao *et al.*, 1987). Vertical differentiation is also displayed by the change in species composition of some transitional genera.

In the Shennongjia region, the basal horizons of the deposits containing assemblage V have an age of ca. 1330 Ma (the U-Pb method, carbonate rocks) (Zhang and Zhang, 1985; Li and Leng, 1987). In the Jixian region, the beds between deposits with the lower and upper stromatolite sets are dated as 1220 - 1190 Ma (the Hongshuizhuang Formation, the K-Ar method, glauconite and a fine illite fraction), and those with the upper stromatolite set, as 1240 - 1020 Ma (the Tieling Formation, the K-Ar method, glauconite). The overlying deposits enclosing assemblage VI yielded a K-Ar date of 1000 - 950 Ma and younger (Zhang, 1985; Xu *et al.*, 1987).

These paleontological and isotopic-geochronological data allow us to correlate assemblage V (Tieling) with the Middle Riphean (Yurmatinian) assemblage of northern Eurasia and to assign the two sets of the assemblage to the Early and Late Yurmatinian, respectively. Thus, there is no reason to further discuss the opinion that assemblage V corresponds to the upper part of the Middle Riphean only (Liang *et al.*, 1985).

The appearance of the two successive sets comprising assemblage V marked the increase in taxonomic diversity of stromatolites in China at the transition from the Late Burzyanian to the Early Yurmatinian, as well as some diversification of species composition at the transition to the Late Yurmatinian (Fig. 3).

The assignment of the two abovelying stromatolite assemblages (assemblages VI and VII) to the adopted scale presents some difficulties that are related to the ambiguous age interpretation of the enclosing deposits in the Chinese literature (see below). The stromatolites of these assemblages have a distinct Late Riphean aspect and are represented mainly by columnar branching buildups. Nearly one-third of the groups and one-fifth of the species of these assemblages are similar to those from the Upper Riphean of northern Eurasia. Among them are the characteristic Late Riphean taxa, such as: *Baicalia rara* Semikh., *Gymnosolen confragosus* Semikh., *G. furcatus* Kom., *G. levis* Kryl., *G. ramsayi* Steinm., *Inzeria tjomusi* Kryl., *Jurusania cylindrica* Kryl., *J. nizvensis* Raab., *Katavia karatavica* Kryl., *Linella ukka* Kryl., *L. simica* Kryl., *Minjaria uralica* Kryl., *Poludia polymorpha* Raab., and others. There are also the long-ranged taxa and numerous endemic genera and species. The differences between assemblages VI and VII are defined by the stratigraphic confinement of mainly the endemic and some of the listed taxa.

There are no stromatolites in the stratotype of the Qingbaikou "System," which corresponds in its time range to the Early Karatavian. However, some elements of assemblage VI appear in the basal horizons of the mentioned "system" in the western Yanshan Mountains. Further to the west in Inner Mongolia (the Beishan and Kuruktag regions and the Chinese Tien Shan), assemblage VI is well represented and replaced by assemblage VII upward the section of the Porgang Group of the Kuruktag region. The succession of assemblages VI and VII was recorded as well from the upper part of the Upper Kunyang Group in the Central Yunnan, the Yangtze platform (Liang *et al.*, 1985). However, these assemblages are best represented in the eastern North China platform, mainly in the Liaonan Group of the Liaodong Peninsula and its correlatives, the Huainan (*s. str.*) and Xu-Huai groups of the adjacent Anhui and Jianxu provinces (Cao and Zhao, 1977; Cao *et al.*, 1985; Xu *et al.*, 1987). It is in these regions that the age of the enclosing deposits is a subject of controversy. Some researchers correlate all these deposits with the lower, pre-tillite part of the Sinian; others relate their lower portion, containing assemblage VI, to the type of the Qingbaikou "System" and the upper part, with assemblage VII, to a younger unit (850 - 700 Ma) referred to the Lower Sinian. This unit is known also as the Liaonan or Huaihe "System," or as an upper part of the Qingbaikou "System" *s. l.* (Sun *et al.*, 1986; Lin *et al.*, 1987; Zhao, 1987; Liang *et al.*, 1987; Xu *et al.*, 1987; Xu, 1987). Yet, some geologists consider the whole Upper Precambrian succession of the Anhui Province as a correlative of the post-tillite Upper Sinian (Zang and Walter, 1992) or even the Lower Cambrian (Vidal and Moczydlowska, 1987).

There are several reasons for the discrepancy in opinions: a difference in the composition of the Upper Precambrian deposits of the Liaodong Peninsula and the Anhui Province from those composing the stratotypes of the Qingbaikou "System" and Sinian; a controversial estimation of the age of the tillites overlying these deposits in the Anhui Province (Lower or Upper Vendian); K-Ar glauconite datings ranging from 600 to 800 Ma; references on occurrences of the supposed medusas (for bibliography see Sun *et al.*, 1986); the presence of carbonaceous sculptured films considered

to be worm remains in the Huainan and Xu-Huai Groups (Sun *et al.*, 1986); the composition of the organic-walled microfossils, particularly the presence of *Baltisphaeridium* sp. and *Ambiguaspora* sp. (Zang and Walter, 1992); and references on the occurrence of *Skiagia* (Vidal and Moczydlowska, 1987).

To evaluate these discrepancies, the following should be taken into consideration. The references on the occurrences of medusas and *Skiagia* were not confirmed (Sun *et al.*, 1986; Zang and Walter, 1992), the nature and time-range of the mentioned films need to be specified, and the representatives of the *Baltisphaeridium* genus were recorded from the Lower Karatavian (Lakhanda) deposits of Siberia (Pyatiletov, 1988). Yet, the stratigraphic uniformity of stromatolite assemblages VI and VII over the whole area of their distribution is unquestionable. This fact and the analysis of all relevant isotopic datings allow us to refine their age estimates.

The shales of the Upper Kunyang Group of the Yunnan, which contain assemblage VI, have a Rb-Sr whole-rock age of 1000 - 950 Ma. The entire group, the upper part of which contains assemblage VII, is unconformably overlain by the basal horizons of the Sinian and is intruded by pre-Sinian granites having the Rb-Sr isochronous age of 861 ± 1 Ma (Huang *et al.*, 1987; Dai and Peng, 1987). The K-Ar age of the basal formation of the stratotype of the Qingbaikou "System" is 956 Ma (illite), and that of the upper formation is 1000 - 800 (glauconite) and 855 (illite) Ma (Zhang and Li, 1982; Zhang and Zhang, 1985). Along with the relatively young K-Ar glauconite datings of 600 - 700 Ma, which can be easily explained by the resetting of isotopic-geochronological systems, the Huainan and Xu-Huai Groups of the Anhui and Jianxu provinces have "older" K-Ar numbers: 976 Ma for the first group and approximately 800 Ma for the second . The isochronous Rb-Sr age of the shales from the Huainan Group was determined as 840 ± 72 Ma (Lin *et al.*, 1987; Xu *et al.*, 1987; Xu, 1987).

These data and partly the composition of stromatolite assemblages VI and VII allow the assignment of these assemblages to the Early Karatavian (1000 - 850 Ma). The co-occurrence of *Chuaria* and *Tawuia* in the enclosing strata (Sun *et al.*, 1986; Zang and Walter, 1992) does not contradict this inference, because they were recently found (Vidal *et al.*, 1993) in deposits referred to the Middle Riphean according to isotopic dating and stromatolite content (Semikhatov and Serebryakov, 1983).

Taxa comprising assemblages VI and VII are responsible for the sharp Early Karatavian peak of generic and, particularly, species diversity of stromatolites in China (Fig. 3). Columnar branching buildups of the Gymnosolenida supergroup dominate the assemblages.

The Late Karatavian stromatolite assemblage is not easily distinguishable in China, because the relevant deposits are mainly clastic and stromatolite-free in the Yangtze platform and are absent from the major parts of the Tarim and North China platforms. The stromatolite-bearing strata of this age have a limited distribution in the southeastern North China platform, in the bordering areas of the Anhui and Jianxu provinces. These strata can be assigned to the Upper Karatavian on the basis of their stratigraphic position, isotopic age, and stromatolite content. They compose the Shouxian Group of negligible thickness, which discordantly lies on the upper horizons of the Xu-Huai Group, contains glauconite that yielded the dates of 650 and 670 Ma, and is disconformably overlain by Lower Cambrian deposits with trilobites (Xu *et al.*, 1987; Zang and Walter, 1992). In the early 1980s, a small shelly fauna was reported from this group (see reference in Zang and Walter, 1992), but this was not confirmed later (Xu *et al.*, 1987). In the Chinese literature, this group is commonly considered as belonging to the Upper Sinian, i.e., the Vendian, because of the mentioned discrepancy in dating the Upper Precambrian deposits of the eastern North China platform (Liang *et al.*, 1985; Sun *et al.*, 1986; Xu *et al.*, 1987).

The microfossils from the Shouxian Group do not allow the specification of its age (Zang and Walter, 1992). As for the stromatolites, they are represented by narrow endemics and local species of genera spread widely in the Upper Riphean and partly in the Vendian (*Acaciella, Anabaria, Aldania, Boxonia, Chihsienella, Gymnosolen, Jacutophyton* (= *Georgina*), *Inzeria, Jurusania, Katavia*, and *Linella*) (Cao *et al.*, 1985; Xu *et al.*, 1987). Regarding these data and the above isotopic age determinations, which define only the minimal age, we assigned the Shouxian Group to the Upper Karatavian. The enclosed stromatolites are much less diverse than those from the Xu-Huai and Huainan Groups.

An impoverished assemblage from the upper part of the Dahuoluoshan Group of the Beishan region, which includes the local species of *Linella* and *Conophyton* and a number of endemics, is also assigned to the Upper Karatavian. The enclosing deposits are separated by a thick siliciclastic member from the underlying horizons of this group, which contain stromatolite assemblage VI, and are unconformably overlain by Lower Sinian tillites (Miao *et al.*, 1984; Zhao *et al.*, 1984). Thus, if the above stratigraphic considerations are correct, the Late Karatavian in China was characterized by a sharp decrease in the abundance and diversity of stromatolites at all taxonomic levels. The decrease would be even more striking if the Shouxian Group belonged to the Vendian or Cambrian, instead of the Upper Karatavian.

The Vendian stromatolites of China are represented by assemblage VIII (Liang *et al.*, 1985), showing some decrease in species diversity while genera diversity remains the same. The assemblage is fully represented only in the Yangtze platform – in the post-tillite Upper Sinian deposits of the Guizhou Province and the Shennongjia region (the Doushantuo Formation) and in the correlative units of the Central Yunnan. Some elements of the assemblage are found in the Xinjiang. It is dominated by local representatives of widespread genera,

such as: *Boxonia, Colleniella, Gongylina, Conophyton, Irregularia, Linella, Paniscollenia*, and others, with subordinate local taxa. The deposits enclosing assemblage VIII were assigned to the Vendian because of their stratigraphic position between the Meishucun Stage of the Lower Cambrian and the Lower Sinian tillites correlative with the Lower Vendian tillites of northern Eurasia (Keller *et al.*, 1984). Studies of the younger stromatolites of China have just begun; there is no evidence for their occurrence in the proven Lower Cambrian deposits.

Thus, the histograms of changes in the taxonomic diversity of the Proterozoic stromatolites of China show three distinct peaks that are synchronous to those established in northern Eurasia (Figs. 2 and 3) but exceed them by an amplitude at the species level. The first peak falls within the second unit of the adopted scale (more precisely, 2.2 - 2.1 Ga); the second one, within the Early Burzyanian; and the highest, third one, which is followed by a sharp diversity decrease, within the Early Karatavian. The noncoincidence in the amplitudes of the synchronous peaks in the northern Eurasia and China histograms is related, to some extent, to the inadequate understanding of taxa volume by Russian and Chinese specialists.

The discussed variations in stromatolite diversity, as noted by Zhu and co-authors (Zhu *et al.*, 1987; Zhu and Chen, 1992), were accompanied by some structural changes in the assemblages – a percentage of members of different supergroup categories (Fig. 3, circle diagrams). The most dynamic categories were columnar buildups, particularly branching ones, and the pre-Middle Riphean time-microstromatites.

In China, the relations between the changes in stromatolite diversity and geological events can be exhibited best by the pre-Riphean buildups which have a limited distribution. A sharp rise in their abundance and diversity 2.3 - 2.0 Ga ago was obviously connected to the radical increase of carbonate rocks in the section of the Hutuo Group and its correlatives and to the enlarged spectrum of environments in which the stromatolite-bearing strata were formed (Zhu *et al.*, 1987). This enlargement is evidenced by the occurrence in the Middle Hutuo of such typical inhabitants of tidal flats as microstromatites along with relatively deep-water morphotypes, including *Conophyton*. However, although the relations are apparent, the increase in stromatolite abundance and diversity in China 2.3 - 2.0 Ga ago cannot be simply a response to regional factors, because the same increase is recorded outside of China as well.

The sharp diversity decrease occurring in China 2.0 - 1.8 Ga ago was a logical result of the wide distribution of molasses of the Luliang orogeny (Goodwin, 1991; Zhu and Chen, 1992), which caused the final consolidation of the North China and Tarim cratons. If present during this time interval, carbonate rocks compose small discrete bodies usually confined to the basal horizons of the epi-Luliang platform cover, which are represented by the Changcheng "System." The insignificant diversity growth 1.8 - 1.65 Ga ago may reflect a wider geographic distribution of carbonate rocks in the upper part of the Changcheng "System" than in the lower one and, possibly, a broader variety of environments, which resulted from the Changcheng transgression.

The considerable increase in the abundance and diversity of the Early Burzyanian stromatolites in China was marked by the appearance of the Mopanyu assemblage, confined to the lower horizons of the predominantly carbonate Jixian "System," which is indicative of a vast sea transgression on the epi-Luliang massives (Chen *et al.*, 1981; Goodwin, 1991). The typical Mopanyu assemblage is confined to the most shallow-water deposits of the Gaoyuzhuang Formation (Song and Gao, 1985) and to the lower part of the first transgressive sequence of the Wumishan Formation. The Chinese geologists recognize four sequences in this formation (Zhu Shixing, per. commun., October, 1992). The deposits are characteristic of a carbonate tidal flat and are analogous to the Upper Burzyanian sediments of northern Siberia, according to our observations. In Siberia, the latter show lower stromatolite diversity than the deposits of an open carbonate platform.

As was demons trated above, the Late Burzyanian decrease in stromatolite diversity, pronounced mostly at the species level, is related to the appearance of assemblage IV (Shanpoling). Its typical representatives – large *Conophyton* and other columnar morphotypes – occur in the upper parts of the mentioned transgressive sequences of the middle part of the Wumishan Formation. The facies confinement of the assemblage is in good accordance with the predominance of conophytonids and the reduced amount of microstromatites. Some diversity increase of the Yurmatinian (Middle Riphean) stromatolites, determined by the appearance of two successive sets of assemblage V (Tieling), proceeded against a background of prevalent environments of the open carbonate platform and a considerably increased Ca/Mg ratio in the Late Yurmatinian strata of some regions (Qin *et al.*, 1985).

The Early Karatavian diversity peak was preceded by a partial cratonization of the Yangtze platform basement and the general rises in the Northern Chinese and Tarim cratons. This peak was developed during the extensive sea transgression which covered some areas elevated in the pre-Karatavian time in the northwest and east of the country (Wang and Qiao, 1984; Chen *et al.*, 1981; Goodwin, 1991) and resulted in the deposition of mainly calcareous sediments (Qin *et al.*, 1985). However, in China, as in northern Eurasia, the great extension of the transgression and the variety of related facies could not be the main, let alone, single cause of the threefold increase in species number and the radical change in generic composition at the transition from the Late Yurmatinian to the Early Karatavian. This inference is supported by the occurrence of the Early Karatavian peak in most superregions rather than by the relationship between the scales of the Late Jixian and Qingbaikou transgressions.

The sharp Late Karatavian drop in stromatolite diversity in the Chinese superregion was undoubtedly caused by the considerable changes in the geological situation after the pre-Sinian Jinning orogeny (860 - 820 Ma ago). Events of this time include: the final cratonization of the Yangtze platform basement, the prevailing siliciclastic sedimentation on this platform in the Late Karatavian, and the elevation of the Tarim platform and most of the North China platform. The shallow-water Late Karatavian basins remained, probably, in limited areas in the southeast of the latter platform. As for the Vendian (Late Sinian) decrease in stromatolite diversity, its relation to the regional events in China shows the same contradictions as those in Eurasia. This decrease proceeded against a background of a vast postglacial transgression and intensive carbonate sedimentation throughout the Yangtze platform in the Late Sinian time. The Late Sinian carbonate strata, similar to the Yudomian of Siberia in all aspects, were deposited in the extensive shallow-water basin, which, although favorable for the formation of stromatolites, was poor in them.

India

In India, Precambrian stromatolite-bearing strata, ranging widely in age from the Late Archean to the Vendian, inclusively, occur very irregularly and sporadically in the time-space matrix. Pre-Riphean stromatolites are reported from the Indian platform. Riphean stromatolites are known from this platform and the foldbelt of the Lesser Himalayas, and the Vendian and Cambrian ones, from the foldbelt only.

The discussed sample of taxa, consisting of 46 formal species and 33 formal genera, does not cover the diversity that is presented in the Indian literature (see bibliography in Stromatolites ..., 1980; Tevari, 1988). Our analysis is based on the revised data published a few years ago (Raaben, 1981, 1985; Raaben and Tevari, 1987) and some later publications (see below). The uncritical treatment of all the taxa reported from the Precambrian of India would be misleading because of the different and sometimes lax approaches of the Indian geologists to stromatolite identification, as well as the ambiguous stratigraphic position of some stromatolite-bearing units. However, even though these units were not analyzed, it was difficult to relate the concrete stromatolite assemblages of the Indian Middle and Late Riphean to the accepted stratigraphic scale.

Archean stromatolites were recorded from the Dharwar craton only (Ventkatachala *et al.*, 1988; Venkatachalapathy *et al.*, 1988; Tevari, 1992), but the data on their composition is controversial. Stromatolites from many local stratigraphic units of the Indian platform belong to the Lower Proterozoic (Kumar, 1976; Stromatolites ..., 1980), but their composition was reliably revealed only for the Aravalli belt of the Rajasthan for the Aravalli and Delhi groups. The Aravalli Group lies unconformably on Archean granites and gneisses, and locally on volcanics having the model Sm-Nd age of 2.6 - 2.3 Ga. The granites intruding into the group have an isochronous Rb-Sr age of 1985 ± 300 and 1900 ± 80 Ma, and the model lead age is ca. 1800 Ma (Crowford and Compston, 1970; Sarkar, 1980; Deb *et al.*, 1989). Thus, this group is referrable to the interval of 2.3 - 2.0 Ga; sometimes its age is estimated as 2.2 - 2.05 Ga (Chauhan and Sisodia, 1988). The Delhi Group is separated from the Aravalli Group by unconformity, involved in tectonic movements with an age of 1.5 Ga (Goodwin, 1991), and intruded by granites with an isochronous Rb-Sr age of 1685 ± 100 Ma (Chaudary *et al.*, 1984). On the basis of these data, we tentatively refer the Delhi Group to the time interval of 2.0 - 1.8 Ga.

The stromatolites of the Aravalli Group are localized in Aravalli phosphorite-bearing limestone. Here, microstromatites of the *Minicolumella* genus, columnar branching *Kanpuria* (their polymorphism caused some misidentifications of several Riphean genera (Banerjee, 1971; Chauhan, 1979; and others)), nonbranching columnar *Conophyton minusculum* Semikh., as well as *Columnacollenia* (= *Masloviella*) and some others (Barman, 1988) were found (Raaben, 1980, 1981). Identifications of the other six genera of the Lower Proterozoic stromatolites from this limestone are questionable (Tevari, 1992), so they were not included in our statistics. The Delhi Group is poor in stromatolites. They are represented by three forms belonging to three genera (Verma and Barman, 1980). In Raaben's opinion, they are parts of single buildups close to *Earaheadia*. In any case, the stromatolites from the Delhi Group are less abundant and diverse than those from the preceding group. Other Lower Proterozoic stromatolites were not taken into account, being inadequately studied or having an uncertain age.

Riphean stromatolites occur in all large basins of the Indian platform (Vindhian, Cuddapah, Pakhal, Kaladgi, and Chattisgarh) and in the Lesser Himalayas. They are in abundance in the Semri Supergroup (Lower Vindhian) of the Vindhian basin but are spread mostly in the lower half of the section – in the lower Semri (Kumar, 1978; Raha and Sastry, 1982; Raaben, 1985; Raaben and Tevari, 1987). The stromatolites constitute two successive assemblages: the first contains *Kussiella kussiensis* Kryl., *Omachtenia, Thyssagetes odontophyes* Vlas., *Colonnella*, and others, and the second is dominated by *Conophyton garganicus* Kor., accompanied by *Colonnella* aff. *discreta* Kom., *Paniscollenia*, and *Gaia*. Some elements of these assemblages occur in the lower horizons of the Cuddapah Supergroup, in the basin of the same name, and in the Purana Group of the Pakhal basin. The assemblages are represented more fully in the lower horizons of the intercorrelated Shali, Gangolihat, and Djammu Big Limestone carbonate formations in the Lesser Himalayas. In addition to *Kussiella* and *Thyssagetes*, the lower assemblage comprises the local species of *Omachtenia* and *Platella*, representatives of the endemic *Rahaella* genus, as well as *Plicatina antiqua* Raab., which is known in Morocco. The upper assemblage consists of *Conophyton cylindricus* Masl., *C. garganicus* Kor., *C. misrae* Kumar, *Conusella regularis* Gol., and a local species of *Colonella* (Kumar

and Tevari, 1977, 1978; Raha, 1980; Valdiya, 1980; Tevari, 1983 a, b; Raha and Das, 1988).

Few isotopic age determinations are available for these stromatolite-bearing deposits. The Rb-Sr dates define the time range of the Semri Group as 1700 - 1150 Ma (Sarkar, 1972). The lower horizons of this group are correlated with the Cuddapah Supergroup, which contains glauconite with the K-Ar age of 1470 ± 60 Ma and lavas dated by the Rb-Sr method as 1370 ± 60 Ma (Crowford and Compston, 1970; Sarkar, 1972, 1980; Chaudary *et al.*, 1984). These dates and the taxonomic composition of the above two assemblages allow their assignment to the Early Riphean and their tentative relation to the Early and Late Burzyanian assemblages of the Urals, respectively (Raaben, 1985). If so, the stromatolites in India were more diverse in the Early Burzyanian than in the Late Burzyanian (Fig. 4).

In the Middle Riphean, carbonates accumulated in limited areas of the Indian platform – in the Vindhian, Cuddapah, and Kaladgi basins (Bozhko, 1984). Stromatolites were reported from the upper horizons of the Lower Semri in the western Vindhian basin only; here, only *Baicalia baicalica* Kryl and *Stratifera* sp. occur (Barman and Verma, 1980). Rare stromatolites from the Upper Semri have not been studied. Middle Riphean stromatolites of the Lesser Himalayas are better known: in the upper half of the Djammu Big Limestone there are some representatives of the *Baicalia* genus, *Anabaria radialis* Kom., nodular *Nucleella* and *Cryptophyton*, and others (Raha, 1980; Raaben and Tevari, 1987). The enclosing beds are underlain by limestones containing a Late Burzyanian stromatolite assemblage with *Conophyton* and are disconformably overlapped by sandstones having a Pb-Pb galenite age of 967 Ma (Raha *et al.*, 1978). This data suggests that the discussed stromatolites should be assigned to the undivided Middle Riphean.

The Upper Riphean deposits, which spread in India more widely than the Middle Riphean ones, are represented mainly by clastic sediments and include only rare stromatolite-bearing units. We took into account the well studied Upper Riphean stromatolites from the Upper Vindhian of the Vindhian basin, the Raipur Group of the Chattisgarh basin, and the Simla Group of the Lesser Himalayas. These units are commonly referred to a single age interval (Raha and Das, 1988).

The Upper Vindhian section begins with the coarse-clastic Kaimur Formation, overlain by the siliciclastic-carbonate Reva Subgroup in which stromatolites including *Inzeria, Tungussia*, and *Baicalia* appear (Prasad, 1980). In addition, *Linella* cf. *ukka* Kryl. is present in the younger Bhander Subgroup. These are the only reliably established taxa, although the general list of stromatolites reported for the Upper Vindhian is somewhat longer. The stromatolite composition is evidence of the Late Riphean age of the Reva and Bhander subgroups. This is in accordance with the K-Ar dates yielded by glauconites from the Kaimur Formation and the underlying Upper Semri – 890 ± 40 and 1080 ± 40 Ma, respectively. The known age estimations for kimberlites [1145 (Rb-Sr) and 1130 (K-Ar) Ma], which intrude into deposits correlative with the Kaimur For-

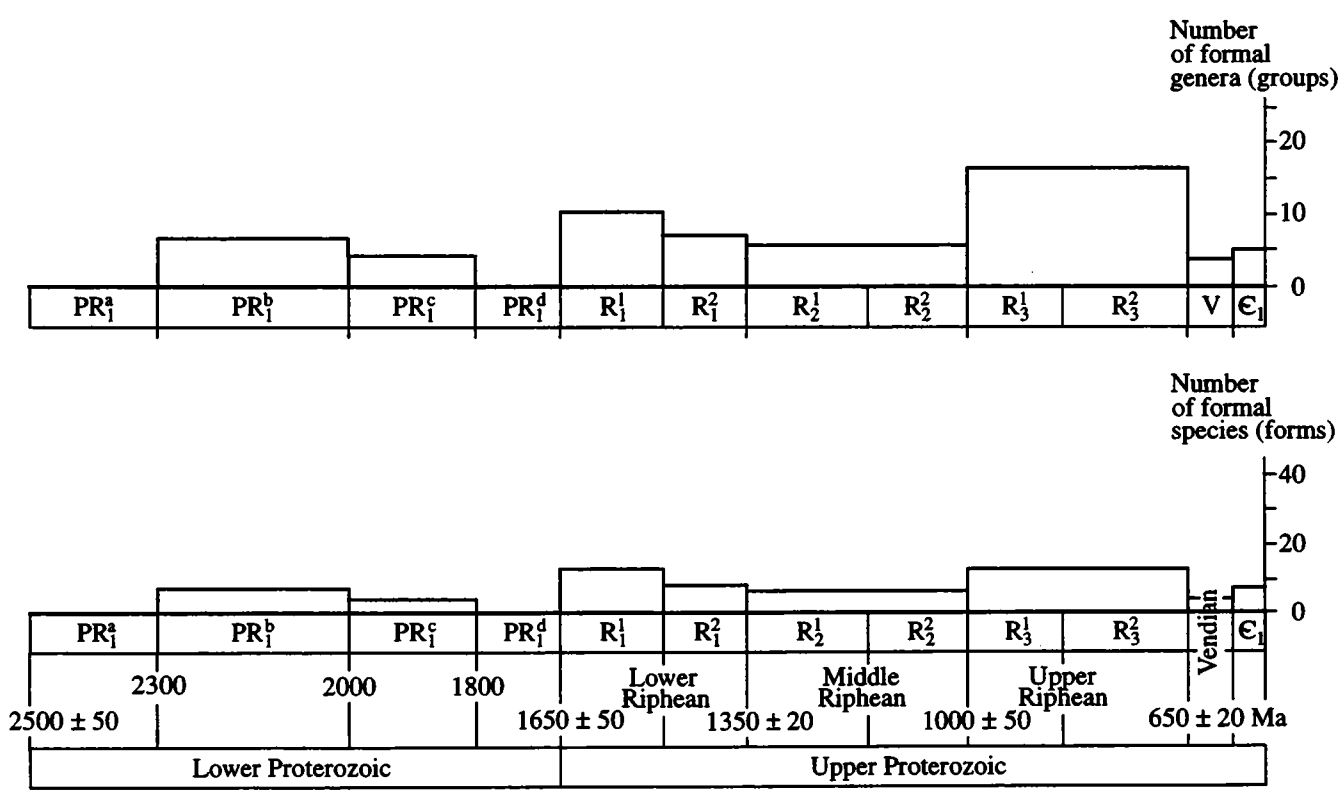

Fig. 4. Dynamics of the taxonomic diversity of the Proterozoic and Early Cambrian stromatolites in India at the levels of formal species (forms) and formal genera (groups). For symbols see Fig. 2.

mation, do not contradict these data (Crowford and Compston, 1970). The kimberlites are unambiguously considered older than the Reva Subgroup enclosing above-considered stromatolites.

The Raipur Group contains a rather rich assemblage of columnar branching forms, including taxa identical to those from the Late Karatavian assemblage of the Urals and assemblages VI and VII of China; these are: *Parmites concrescens* Raab., *Minjaria, Gymnosolen*, and *Inzeria*. There are also present endemic *Tungussia indiana* Raab. with a typical Upper Riphean microstructure *Fibrosida* (Raaben and Zabrodin, 1972; Bhattacharia, 1975; Jairman and Benerjee, 1980), *Conophyton*, and rare *Platella*. The occurrence of these stromatolites supports the Late Riphean age of the Raipur Group (Schnitzer, 1977; Sarkar, 1980; Raha and Das, 1988); however, it is inconsistent with the correlation of this group to the Lower Vindhian, as noted in the Indian literature (Jairman and Banerjee, 1980).

The Simla Group lies on the Shali Formation, the thick lower part of which is correlated with the Djammu Big Limestone (Valdiya, 1980). The group is composed of siliciclastic deposits with large lenses of stromatolite limestones. The lowermost lens, sometimes referred to the upper part of the Shali Formation, encloses *Parmites concrescens* Raab. (Raaben and Tevari, 1987; Sinha and Raaben, 1979), and the other lenses contain *Jurusania himalaica* Sinha and some other taxa (Sinha, 1977). These finds allowed assignment of the Simla Group to the Upper Riphean–Vendian (Sinha, 1977; Raha and Das, 1988). Presently, the minimum age limit of this group is determined by its position under the Blaini tillites, which belong to the Lower Vendian glaciohorizon (Tevari, 1992). The age relationships between the Late Riphean stromatolites of the three discussed regions are unclear. Therefore, all of them are referred to the undivided Upper Riphean. Thus, the latter differs from the Middle Riphean in its elevated stromatolite diversity.

Vendian stromatolites of India are known from the Krol Group of the Lesser Himalayas, which lies between the Blaini tillites and the Lower Cambrian deposits. Their taxonomic composition is under study (Tevari, 1992). According to Raaben's observation, it consists of three or four taxa. If so, the drop in diversity of the Vendian stromatolites can be logically inferred from their local distribution in India. Being locally distributed also, the Early Cambrian stromatolites from the Krol Group (Tevari, 1992) are more diverse than the Vendian ones.

The performed analysis allows us to make the following conclusions:

(1) In each of the three Eurasian superregions, the dynamics of stromatolite diversity show the same trends at species and generic levels. In essence, the histograms of variations in species abundance differ from those compiled for genera in the greater amplitude of the variations; the sign for the latter is the same (Figs. 2 - 4).

(2) There is a close similarity between the histograms of changes in the most representative samples of stromatolite taxa from northern Eurasia and China. They are similar in their three-modal character and the positions of the peaks and depressions. The depressions occurred at the very beginning of the Proterozoic (2.5 - 2.3 Ga), at the end of the Proterozoic (Vendian and, partly, the Late Karatavian), at the end of the Early Proterozoic, and in the Late Burzyanian (Figs. 2 and 3). The peaks were recorded in the middle of the Early Proterozoic (2.3 - 2.0 Ga), the Early Burzyanian, and the Early Karatavian. In northern Eurasia, the third, Early Karatavian peak exceeds the other two at the species level and is similar to them at the generic level. In China, all three peaks of species diversity are eventually similar, although the Early Karatavian peak remains higher. The highest peak of genera diversity in China rather unexpectedly appears to be the Early Proterozoic one, which is defined by the stratigraphic position of stromatolites of the relatively small region.

(3) Against the background of quantitative variations, the successive stromatolite assemblages of northern Eurasia and China exhibit directed structural changes – redistribution of the relative roles of supergeneric taxa in their composition. The general trend of these changes appears to be very similar in both superregions during the Riphean (see circle diagrams in Figs. 2 and 3). In China, the pattern is complicated by the noticeably increased role of microstromatite taxa (*Asperia*–form stromatolites) at the end of the Early Proterozoic and, especially, in the Early Burzyanian.

(4) The histogram compiled for the Proterozoic stromatolites of India is much less representative, being based on selected data from the well-studied areas of this relatively small superregion. Therefore, the pattern of the histogram (Fig. 4) should not be absolutized, especially since the diversity differences are rather small and in some cases comparable to statistical noise. Nevertheless, this histogram shows a similarity with the more representative north Eurasian and Chinese histograms in the increased stromatolite diversity in the middle of the Early Proterozoic (2.3 - 2.0 Ga), in the Early Burzyanian, and at the transition from the Yurmatinian to the Karatavian, as well as decreased diversity in the Vendian. However, the histograms for India differ greatly from those for China and northern Eurasia in the Middle Riphean part, which demonstrates impoverished stromatolite composition. This peculiarity could be related to the poor knowledge of the Indian material if it was not repeated in the histograms for other superregions belonging to the Gondwana group.

(5) Analysis of the geological data on northern Eurasia, China, and India reveals that the variations in diversity of pre-Vendian stromatolites were related to vast transgressions (maxima of the diversity), orogenic epoches, and regressions (minima of the diversity). The Vendian stromatolite decline was rather specific and proceeded in northern Eurasia and southern China during wide postglacial transgressions.

(6) Similar trends in the dynamics of diversity of the Proterozoic stromatolites in these superregions suggest

control by events and processes, which were common, for at least the majority of Eurasia.

ACKNOWLEDGMENTS

The authors sincerely thank S.M. Awramik, M.B. Bursin, Zhu Shixing, A.H. Knoll, Liang Yuzuo, and M.A. Fedonkin for their fruitful discussion of the problems under consideration in this article and their critical comments. This study was supported by the Russian Foundation of Fundamental Research (project no. 93-05-09403).

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"Gigantism" in Precambrian Microfossils and its Biostratigraphic Significance

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Received May 31, 1993; in final form, July 17, 1993

Abstract – The analysis of statistical data on the age-related distribution of some genera of Precambrian microfossils suggests that at certain periods of development there appears a stage characterized by "gigantic" forms. In different genera, this stage is assigned to different age intervals, which are much shorter than the entire time-range of the genus itself. This phenomenon is of scientific and, in particular, biostratigraphic significance for the Precambrian, as such forms allow for more precise dating and correlation of host deposits.

At first, the very concept of "microfossil gigantism" seems illogical. Nevertheless, it is reasonable to put forward such a problem. Gigantic forms are defined not by their absolute size, but by their size with respect to other specimens and objects of the same category or class. As was vividly shown by Jonathan Swift, Gulliver seemed to be a giant in the country of Lilliputians, whereas in the country of giants, he himself appeared a Lilliputian. Examples of gigantic forms in various fauna and flora groups at certain stages of their development are well-known in paleontology.

The general tendency toward size increase during phylogenetic evolution in various taxonomic groups was pointed out as long ago as the end of the 19th century by the then outstanding paleontologist E. Kop (Raup and Stanley, 1971). This phenomenon received the name "Kop's rule" or "Kop's law." Later, Sh. Depere (1921) qualified it as "the law of increasing growth in phylogenetic branches." He believed that "by its generality it is one of the most interesting and significant laws established by works of recent paleontologists" (*ibid*, p. 152). Depere cited several examples of gigantic forms in certain animal genera, including unicellular ones (foraminifers of the *Orbitolina* genus). Numerous examples of gigantic forms of plants can be found in *Paleobotanics* by A.N. Krishtofovich (1957). Researchers explain the increase in plant and animal size with different causes, such as: a change in climatic and other conditions, protective functions generated by organisms, the narrower specialization of organisms, and other causes. These issues are most comprehensively described in the works of I.I. Shmal'gauzen (1968), A.N. Krishtofovich (1957), D.M. Raup and S.M. Stanley (1971), and others.

Unfortunately, in recent decades the issue of gigantism in plant and animal life no longer seems to attract the attention of Russian paleontologists. In modern reference books and manuals the term "gigantism" is absent, and Kop's and Depere's Rules (or Laws), if mentioned at all, are qualified as empiric ones, not reflecting the natural laws of flora and fauna evolution. In this connection, it is pertinent to cite V.I. Vernadskii, who said: "I believe that empiric generalizations are much more valid in geology than scientific hypotheses even if they are so far not understood, for they are found to be in agreement with reality and so should be accepted as scientific fact" (Vernadskii, 1965, p. 225). We think that on no account should the problem of gigantism be disregarded in paleontological investigations, even if generalizations are empiric in character.

The establishment of gigantism stages in the development of Precambrian microorganisms would be of both biological and biostratigraphic significance. The period of gigantic form development embraces far shorter intervals than the existence of a single group as a whole. With regard to the generally accepted opinion of the slow morphological evolution of Precambrian organisms, the identification of giant forms among separate groups would enable more precise and reliable dating of host deposits and their correlation in distant sections. The establishment of gigantism stages in the development of separate genera would be of essential importance. However, in solving this problem, we run into certain obstacles.

The expression "gigantic microfossils" occurs in the literature on Precambrian micropaleontology, though it is used only to denote large-sized acritarch, which are regarded as highly developed eukaryotic organisms. As for prokaryotic microorganisms, many foreign researchers are of the opinion (shared by some Russian micropaleontologists) that these microorganisms are the most conservative and are represented by smaller forms than eukaryotic ones.

Nevertheless, the analysis of the available data suggests that there were gigantic forms among Precambrian prokaryotes, which occurred at certain stages in the development of some genera and at different stratigraphic intervals. However, the identification of stages with gigantic forms is complicated by the fact that the size of microfossils is regarded as a taxonomic feature

of genus rank, and large and small forms are described under different genera names, though in structure they are completely identical. Examples of the differentiation of such genera will be given below.

The predominant part, if not the overwhelming majority, of Precambrian microfossils in dark, fine-grained cherts in carbonate successions are compared with recent cyanobacteria or blue-green algae. It is remarkable that certain genera of these microorganisms, both coccoid and filamentous, show the presence of "gigantic" forms in addition to small ones. Thus, according to Gollerbakh *et al.* (1953), different species of the *Synechococcus* Näg genus are represented by solitary ellipsoidal cells, their width ranging from 1.4 - 2.0 to 39 - 52 μm. In colonial coccoids of the genus *Gloeocapsa* (Kütz) Hollerb., the cell diameter of various species ranges from 2 - 3 to 54 - 58 μm, with sheaths up to 60 - 70 μm. Among filamentous blue-green algae, the width of trichomas (equal to the cell width) in various species of the genera *Oscillatoria* Vauch. ranges from 1.0 - 1.2 to 25 - 50 μm and even to 70 μm. In the genus *Lyngbya* Ag., the width varies from 1.0 - 1.5 to 20 - 40 μm, sometimes reaching 60 - 80 μm. The size of large forms in all the cited examples are more than one order of magnitude larger than the size of small forms, and those of the filamentous algae are almost two orders of magnitude larger, i.e. gigantism is clearly observed among the considered genera. Regretfully, we do not know the literature describing the causes and time of appearance of those species represented by small and gigantic forms.

Among Precambrian microfossils, gigantism is most clearly observed in the genera *Obruchevella* Reitlinger and *Eosynechococcus* Hofmann, though mostly in the former. *Obruchevella* is characterized by a peculiar structure and represented by isolated tubular filaments with different diameters twisted to form a regular cylindric spiral. This clearly differentiates them from other filamentous microfossils and makes it possible to identify them both in chert thin sections and macerations of terrigenous rocks. Currently, all researchers recognize the similarity between *Obruchevella* and recent blue-green algae of the genus *Spirulina* Turp.

Obruchevella remains have been found widely distributed in many regions (Fig. 1). They are known from many areas of the Siberian platform and its folded framework (Patom uplift, Irkutsk amphitheatre, southeastern part of the Eastern Sayan, Enisei Ridge, and elsewhere), as well as in Kuznetsk Alatau, Tuva, southern Kazakhstan, and the Urals. In addition to northern Eurasia, they are determined in Mongolia, the northern part of Saudi Arabia, certain Chinese provinces, eastern Alaska, Spitsbergen, and Greenland. *Obruchevella* is highly diversified; more than 15 species have been distinguished and described. Detailed lists of localities with *Obruchevella* species, their related spiral-like forms, and their stratigraphic confinement are given in recently published articles by V.K. Golovenok (1989) and K. Mankiewicz (1992).

Wide geographical spread and high species diversity make *Obruchevella* a favorable subject for investigating the gigantism phenomenon in certain genera of Precambrian microfossils and for evaluating its biostratigraphic significance. However, before considering the indicated problems, we should touch upon the spiral-like microfossils identified by researchers under other genera names. Among them there are microfossils assigned to the genera *Heliconema* Schopf, *Avictuspirulina* Strother *et al.*, and *Volyniella* (Schep.) Asseeva. Their similarity (or even identity) with *Obruchevella* was indicated earlier (Golovenok, 1989). It is noteworthy that the authors compared the first two genera with recent *Spirulina*. In China, India, and Greenland, in the Precambrian–Cambrian boundary deposits (in the Meishucun, Lower Tal, and Portfield formations, respectively), large spiral-like microfossils are identified under the genera names *Spirellus*, *Glomerula*, and *Jangispirellus*. J. Peel (1988) indicated their similarity and relation to *Obruchevella*; however, with regard to their large size, he assigned them to two separate genera – *Spirellus* (*Glomerula* inclusive) and *Jangispirellus*.

The taxonomy of all the above-discussed genera of spirally twisted microfossils requires thorough revision. We believe that they most likely belong to a single genus, which, in accordance with priority rule, should bear the name *Obruchevella* Reitlinger; the other names should be regarded as junior synonyms. The sizes of the tubes and spirals, and the extension of the latter, are species features and cannot serve as a basis for identifying new genera. The presence or absence of membranes inside tubular filaments cannot be regarded as such either. In various species of the recent genus *Spirulina* Turp., with which the microfossils are compared, the cell membranes inside trichomas are not always visible (Gollerbach *et al.*, 1953).

Spirulina representatives are not the only ones to have a spiral-like form. This form is also characteristic of recent blue-green algae of the genus *Lyngbya* Ag.; but unlike *Spirulina*, *Lyngbya* Ag. is characterized by a well-expressed cellular structure of trichomas and by the presence of a sheath. Similar microfossils of the genus *Palaeolyngbya* Schopf are known in the Precambrian; however, no spiral-like forms were identified among them.

Of particular interest is the analysis of the size and age distribution of *Obruchevella* and other spiral-like microfossils (Fig. 2 A, Plate, Figs. 1 - 3). As of yet, they have not been discovered in deposits more ancient than the Late Riphean. Late Riphean forms are always very small. The diameter of the tubes is only several microns, and the outer diameter of the spirals ranges from 4 - 5 to 18 - 20 μm. The sizes of the microfossils under consideration substantially increased in the Vendian and "gigantic" forms appeared among them. Recently, we have established and described two such "gigantic" *Obruchevella* species from the Vendian deposits of the Buton Formation in the Central Urals (Golovenok *et al.*, 1989) and one from the Parshino

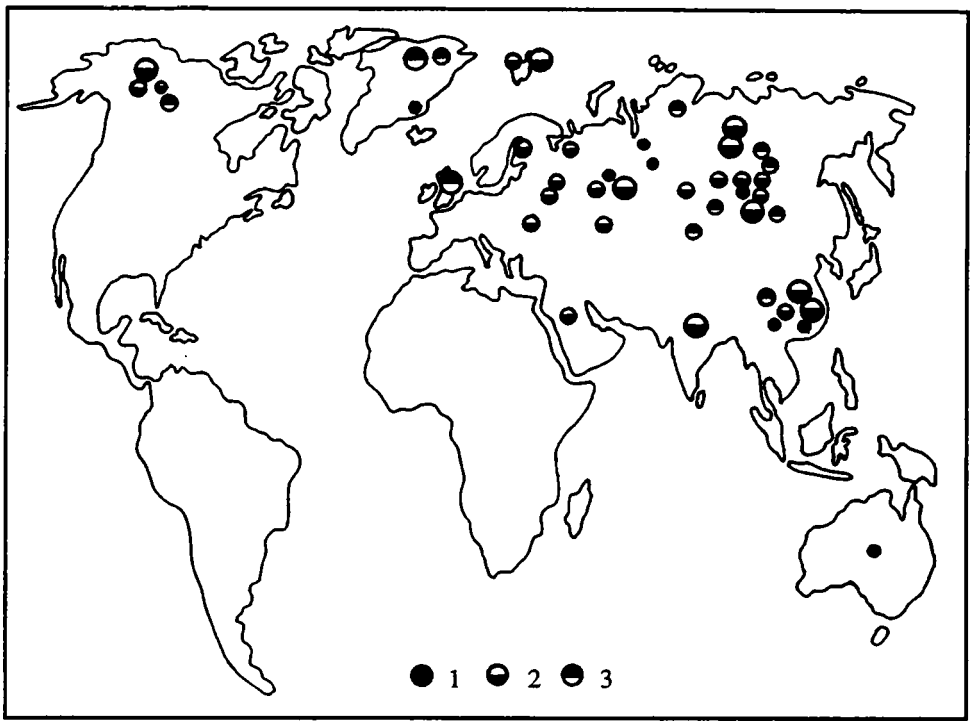

Fig. 1. Geographic and stratigraphic distribution of *Obruchevella* Reitlinger.
1 – Upper Riphean, 2 – Vendian, 3 – Lower Cambrian. The symbols show relative dimensions of forms: small, medium, and large.

Formation of the Nepa–Botuoba structure of the Siberian platform (Golovenok *et al.*, 1990). The diameter of *Obruchevella* tubes in the Buton Formation reaches 45 - 55 μm, while that of spirals – 275 - 360 μm. Similar "gigantic" *Spirellus* and *Jangispirellus* representatives were described in detail from the Precambrian–Cambrian boundary deposits in the above-mentioned article by J. Peel (1988). Only *Obruchevella* forms were described from younger Cambrian deposits. Incidentally, they were smaller in size than those mentioned above. The data reported by E.A. Reitlinger (1959) and some other researchers indicate that the diameter of tubes of Cambrian *Obruchevella*, including type species, is 12 - 18 μm, and the diameter of spirals varies from 32 - 36 to 58 - 68 μm. "Gigantic" forms of *Obruchevella* were not recorded in Cambrian and other Paleozoic deposits.

The genus *Eosynechococcus* Hofman is another characteristic genus of Precambrian microfossils with a pronounced "gigantism" stage. Microfossils of this genus are represented by ellipsoidal and elongated-ellipsoidal to rod-like forms with rounded edges. Cell sizes of different species vary over a wide range (Plate 1, Figs. 6 - 8). Their width ranges from several microns to several dozens of microns, and the width–length ratio in different species varies from 1 : 2 to 1 : 6. Large cells commonly occur one at a time, more rarely in pairs, whereas small ones may be assembled in loose or compact groups (colonies). This genus, consisting of three species, was identified for the first time in cherts of the Early Proterozoic Belcher Supergroup in Canada (Hofmann, 1976). Later, *Eosynechococcus* representatives, and forms similar to or identical with them were established in cherts from Precambrian deposits in many regions of the world (Greenland, Spitsbergen, China, India, and Australia). In the territory of the former USSR, they are known from Riphean deposits of the Anabar and Turukhansk uplifts, Enisei Ridge, the southern Urals, and southern Kazakhstan. The time range of their development extended from the Early Proterozoic to the Late Riphean, inclusive. They have not yet been identified in the Vendian. At the present time, more than 10 species of the genus have been described.

The Middle Riphean is characterized by the appearance of large forms, being almost one order of magnitude greater than small ones, as described by Hofmann (Fig. 2 B). So far, no such "gigantic" *Eosynechococcus* forms have been recorded in either older or younger deposits. As in the case of *Obruchevella*, certain researchers regard the dimensions of *Eosynechococcus* cells as genus criterion, and large forms are distinguished under different genera names. Thus, Horodyski and Donaldson (1980) singled out such microfossils from the Middle Riphean (1175 - 1225 Ma) deposits of the Dismal Lakes Group (northwestern Canada) as the separate genus *Archaeoellipsoides* and attributed them to a single species – *A. grandis*. However, according to the authors, the dimension range of these microfossils varies over wide limits: the width of ellipsoids, from 5

to 33 μm, and their length, from 9 to 92 μm. The researchers noted that smaller *Archaeoellipsoides* forms are similar to *Eosynechococcus grandis* Hofmann from the Belcher Supergroup, but no clusters or scattered aggregates, as were described for *Eosynechococcus*, occurred among them. This served as the basis for distinguishing a separate genus.

In this connection, it should be noted that Hofmann identified as the type species (*E. moorei*) of the genus *Eosynechococcus* the small rod-like cells, forming irregular clusters (colonies) and occurring as solitary representatives or 3 - 4-cell chains. This choice of a type species seems most inappropriate, as the recent genus *Synechococcus* Näg., to which he compared microfossils from the Belcher Group, does not form cell accumulations (colonies) (Gollerbach *et al.*, 1953). Cells of other *Eosynechcoccus* species identified by Hofmann, unlike the type species, occur one by one or in the form of 2 - 3-cell chains. In morphology, cell size, and colonial habit, microfossils of the type species are similar to recent blue-green algae of the genus *Gloeothece* Häg., which was previously noted by S. Golubic, S. Kempbell, and others.

In the cherts from the Kotuikan Formation of the Anabar uplift, we described ellipsoidal, occasionally rod-like microfossils of the genus *Eosynechococcus* Hofmann, among them very large ones, grouped into several species (Golovenok and Belova, 1984). Isolated specimens of these microfossils, represented graphically according to their size, demonstrate that the transition from typical *Eosynechococcus* Hofmann (with *E. medius* and *E. grandis* interpreted as such) occurs gradually along a single line (Golovenok and Belova, 1984, Fig. 1a). According to Horodyski and Donaldson (1980, Fig. 1b), the same size distribution is observed in *Archaeoellipsoides*. In fact, an overlap of point locations is observed in both diagrams. We believe that there is no sufficient evidence to distinguish the independent genus *Archaeoellipsoides*. All microfossils described under this name should be attributed to the genus *Eosynechococcus* Hofmann.

Following the identification of the separate genus *Archaeoellipsoides*, various opinions appeared on the biological affinity of these microfossils. Horodyski and Donaldson believed that this problem still remained obscure. However, the large dimensions of some of the microfossils led the authors to assume that these might be either the outer sheaths of colonial microorganisms with completely degraded inner contents, or the isolated spores or sporangia of blue-green algae, or cysts of eukaryotic algae. The latter supposition is shared by some Russian micropaleontologists. All the suppositions are based exclusively on the sizes of the considered microfossils and are supported by the belief that prokaryotic microfossils cannot have large cells, though, as we already mentioned, that is far from being true in reality.

Gloeodiniopsis Schopf are widespread geographically and are characterized by wide a time range. They are represented by solitary spheroidal cells with thick walls and distinct multilayered sheaths (Plate 1, Figs. 3 - 5). The latter characteristic allows differentiation of

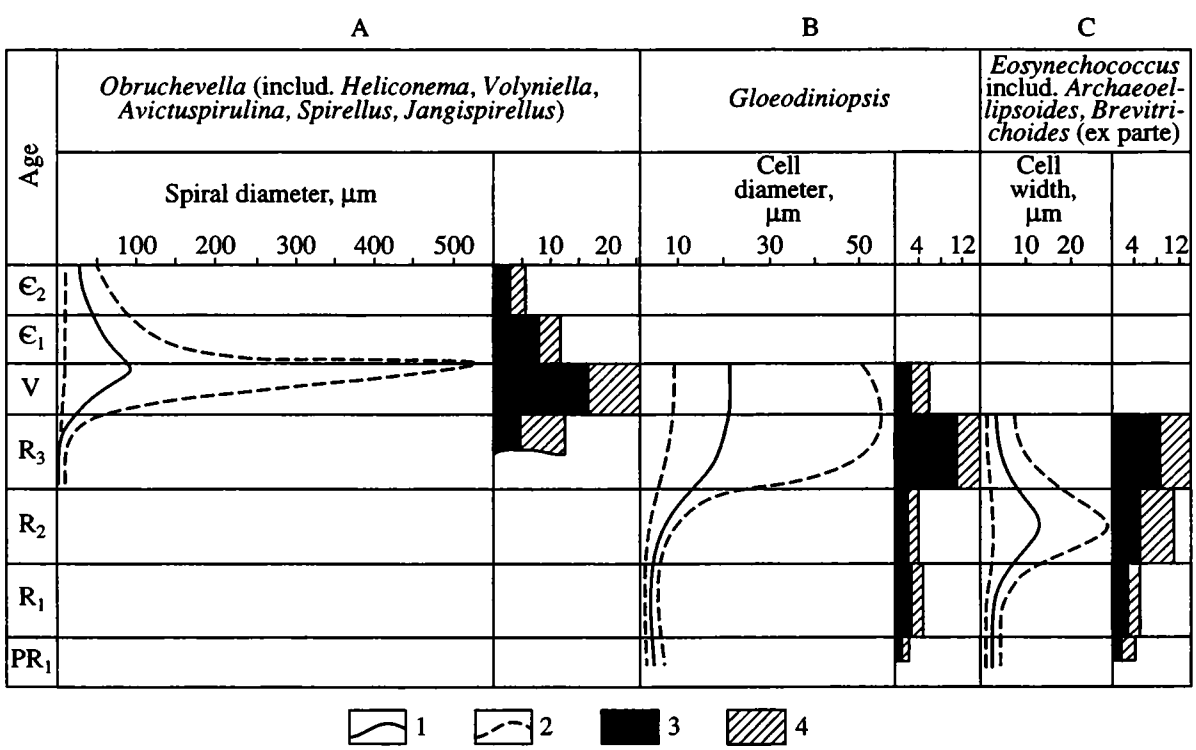

Fig. 2. Size variation in some Precambrian microfossil genera in time.
1 – medium size, 2 – size range limits, 3 – number of localities, 4 – number of species.

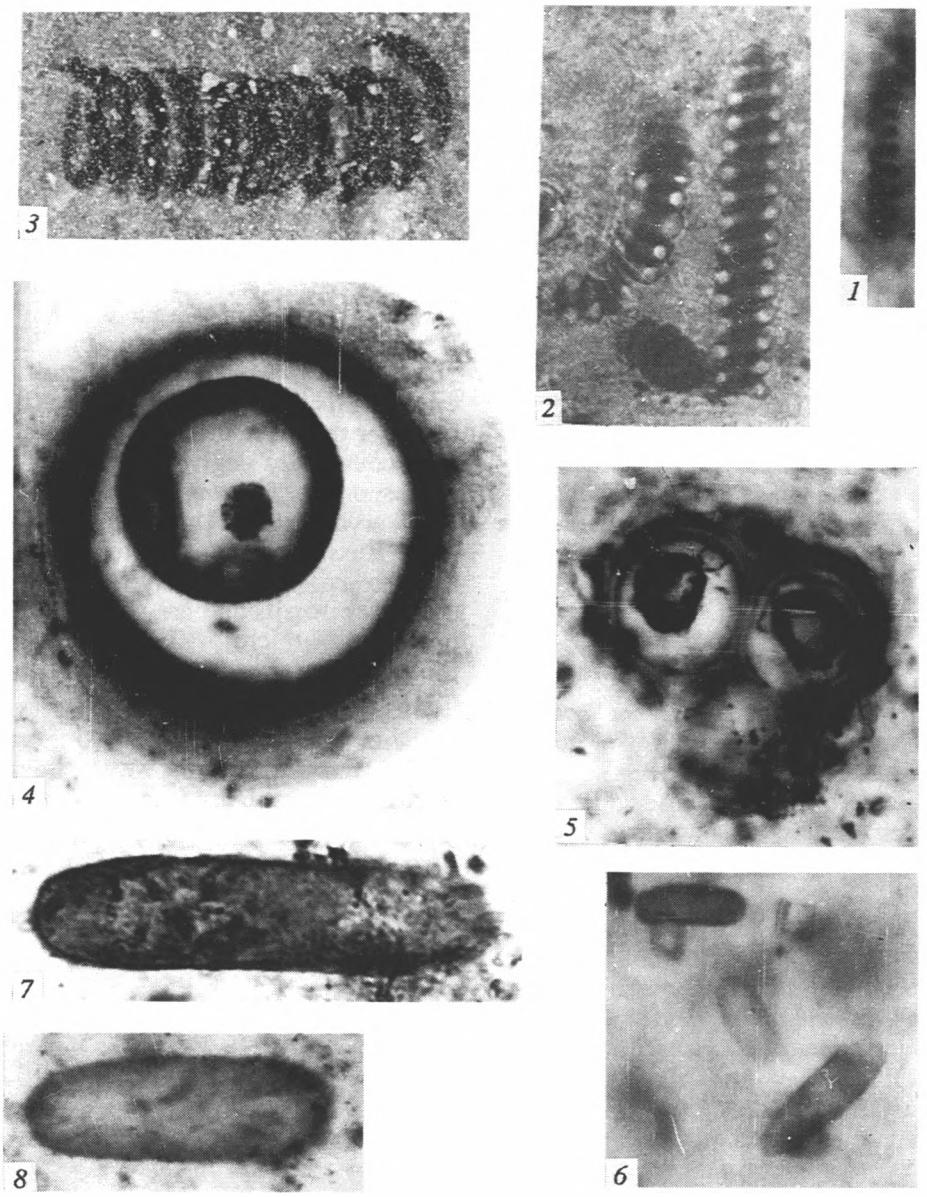

Plate 1. Some silicified microfossils from the Riphean of Siberia.
1 – *Obruchevella pusilla* Gol. et Bel.; ×1000; Valyukhta Formation, Upper Riphean, Patom uplift.
2 – *Obruchevella parva* Reitlinger; ×250; Tinnaya Formation, Vendian, Patom uplift.
3 – *Obruchevella* cf. *gigantea* Gol. et Bel.; ×130; Parshinskaya Formation, Vendian, Siberian platform.
4 – *Gleodiniopsis magna* Nyberg et Schopf; ×1000; Sukhaya Tunguska Formation, Upper Riphean, Turukhansk uplift.
5 – *Gloeodiniopsis lamellosa* (Schopf) Knoll et Golubic; ×800, Turukhansk uplift.
6 – *Eosynechococcus grandis* Hofmann; ×600; Kotuikan Formation, Lower Riphean, Anabar uplift.
7 – *Eosynechococcus giganteus* Gol., et Bel.; ×600, *ibid.*
8 – *Eosynechococcus major* Gol. et Bel.; ×600, *ibid.*

Gloeodiniopsis from all other single coccoid microfossils and its adequate identification. Approximately 10 species have been identified in the genus. Occasional finds of this genus are known from the Early Proterozoic in Canada (Horodyski and Donaldson, 1980), though they are most typical of Riphean and Vendian deposits. It is noteworthy that *Gloeodiniopsis* was represented by relatively small (10 - 12 μm) forms in the Early Proterozoic and the Early and Middle Riphean (Fig. 2 B). *G. magna* Nyberg et Schopf appeared in the Late Riphean; their cell diameter was 30 - 36 μm, along with broad sheaths – 50 - 56 μm (Plate 1, Fig. 4). This species was first identified in the cherts of the Min'yar Formation in the southern Urals (Nyberg and

Schopf, 1984). We revealed it in the Sukhaya Tunguska Formation of the Turukhansk uplift (Plate 1, Fig. 4). Microfossil dimensions of this genus are smaller in Vendian deposits. Thus, *Gloeodiniopsis* also shows a tendency for the appearance of "gigantic" forms at a certain stage of development.

Large-sized representatives of certain microfossil species are characteristic of the genus *Oscillatoriopsis* Schopf; they are confined to the Middle Riphean deposits. The width of their trichomas may reach 50 - 55 μm and more (in the type species it is 3.7 μm on average). The width of trichomas in Early Riphean, Late Riphean, and Vendian forms is substantially less. More than 20 species are distinguished in the considered genus, and approximately 10 forms are described under an open nomenclature. The species diversity of this genus seems to be even higher, as microfossils similar to it in structure and dimensions are described under other genera names. In this connection, we should emphasize that presently, more than 40 genera have been identified among Precambrian microfossils in the family Oscillatoriaceae (Kirchn.) Elenkin, though no more than 10 genera are known among its recent representatives. Such an abundance of genera in Precambrian microfossils assigned to this family can be explained by the insufficiently rigorous and justified delimitation of generic (evidently, species as well) taxa in its composition. For this reason, it is impossible to reveal stages with "gigantic" forms in the evolution of certain Precambrian Oscillatoriacean genera without a thorough revision of the taxonomic composition of the microfossils considered.

Thus, the size analysis of certain characteristic groups of Precambrian microfossils suggests that the stage of gigantic forms occurred in the development of several genera of Precambrian microorganisms. The causes for the appearance of gigantic forms remain unclear. This problem requires special, primarily ecological research. From the viewpoint of applied biostratigraphy, it is significant that gigantic forms of different genera appear in different age intervals: *Eosynechococcus* – in the Middle Riphean, *Gloeodiniopsis* – in the Late Riphean–Vendian, and *Obruchevella* – in the Vendian–earliest Cambrian. If all these conclusions are confirmed in the course of further investigations, we shall have a valuable paleontological tool for Precambrian biostratigraphy.

We are well aware that the conclusions presented in this article are empirical and based on limited statistical material (though it is quite extensive for *Obruchevella*) and so must be further discussed. Nevertheless, we would like to draw the attention of researchers to the occurrence of gigantic forms in the development of certain ancient microorganisms. To confirm or rule out these conclusions, much new evidence is required, along with a thorough analysis of the available data and a critical review of the taxonomic composition of Precambrian microfossils.

Reviewer A.Yu. Rozanov

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Stratigraphy and Geological Correlation, Vol. 2, No. 6, 1994, pp. 520 - 541. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 2, No. 6, 1994, pp. 40 - 63.
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Global Rhythms of the Phanerozoic Endogenic Activity of the Earth

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Received September 20, 1993; in final form, December 15, 1993

Abstract – During the Phanerozoic, tectonic deformations, granitization, metamorphism, and volcanic processes continuously manifested themselves in one region of the Earth or another, though with varied intensities. This study revealed that generally synchronous peaks of deformation activity, island-arc volcanism, and regional metamorphism were often coeval with those of granitization and volcanic activity along continental margins. Concepts of global epochs of tectogenesis and four tectonic cycles (the Caledonian, Hercynian, Kimmerian, and Alpine) of the Phanerozoic proved to be realistic. These cycles were parts of megacycles (the Wilson's cycles), which lasted for approximately 600 Ma. The megacycles covered the time span from the breakup of Pangea until the moment when its fragments, with newly accreted continental crust, joined to form a new Pangea. We recognized 16 Phanerozoic epochs of tectogenesis, nearly always separated by intervals of 35 - 40 Ma. There were four periods with the most prominent endogenic activity of the Earth, which were terminal in the tectonic cycles. They corresponded to the Late Silurian, Late Permian, Late Jurassic, and to the final stage of the Miocene–Pliocene. Their intervals were not constant, varying from 120 to 150 Ma. The last of these periods belongs to the Alpine cycle, which has already been lasting approximately 140 Ma up to the present terminal paroxysm.

INTRODUCTION

The Phanerozoic, which lasted for approximately 530 - 550 Ma, is the best studied and most well-known part of the Earth's history. The Phanerozoic tectonic evolution of the planet was motivated first by continental coupling into the supercontinent Pangea, and then by its breakup at the beginning of the Mesozoic. These events reveal processes occurring in the Earth's core and mantle. In this paper, we intend to briefly review the tectonic history of the Earth during the Phanerozoic; to correlate manifestations of such endogenic processes as granitization, metamorphism, tectonic deformations, formation of ophiolites, island-arc volcanism, trap magmatism, and rifting; and to analyze global secular variations in the intensity of these processes. Our goal is to define the reality of Phanerozoic global rhythms of the endogenic activity of the planet and to reveal main tectonic epochs and eras, the periodicity of which reflects the most general regularities in the evolution of the Earth's tectonosphere.

The history of recognizing cycles in the evolution of the Earth crust is longer than 150 years. Many researchers were, to some extent, concerned with problems of global rhythms of tectonic processes. The "canon of orogenic phases" (in fact, the phases of folding) by German geologist Stille, first edited in a general form in 1924, is the most well known. Earlier, at the end of the 19th century, there were published works by French geologist Bertrand, who considered problems of tectonic cycles. Using data on studied sequences of characteristic sedimentary formations, Bertrand showed that the Alpine geosyncline went through definite evolutionary stages during the Mesozoic and Cenozoic. He also discovered that the same stages are recognizable in the evolution of Paleozoic fold systems in Europe and in the Late Precambrian fold belts of North America. By these means, the Huronian, Caledonian, Hercynian (Variscan), and Alpine tectonic cycles were distinguished.

Later, both the concept of orogenic phases (phases of folding) and the notion of tectonic cycles met serious objections. Ideas on the periodic, episodic, and intermittent characteristics of tectogenesis were sometimes replaced by alternate viewpoints based on the continual, or even chaotic occurrences of tectonic processes (Pushcharovskii, 1993). The objections are understandable, because phases of tectogenesis are sufficiently synchronous only for large structural elements (systems). As was shown by our correlations at the global rank (Khain and Seslavinskii, 1990, 1992), such phases are actually concentrated within certain intervals of geological history (e.g., in the Late Silurian, Late Jurassic, etc.).

The notion of tectonic cycles was and is under criticism due to the following: (1) the term "cycle" suggests a strict recurrence of events that is absent in reality and (2) investigations within particular mobil belts and systems put in the forefront not a coincidence between accumulation periods of volcano-sedimentary complexes and periods of orogenesis and folding, but the

definite trends in evolution leading to a general structural complication.

Although such objections are correct, we think they are insufficient to negate cycles. From a dialectic standpoint, the cycles need not be identical. On the contrary, they are parts of larger cycles, because the evolution does not follow a closed circle. Respectively, each subsequent cycle regularly differs from the preceding one, and the recurrence and directivity of evolution are combined. The recurrence is evident only in the most general features, such as the alternating environments of dominant extension and compression, or the repeating replacement of ophiolites by island-arc and continental volcanic complexes, as well as in periodic changes either from abyssal to shelf and terrestrial sedimentation or from basic to acid magmatism. The collision processes associated with intensive deformations, regional metamorphism, and areal expansions of orogenic regimes are also periodic and usually confined to the end of a cycle.

Some of these events may not be realized, or they may be manifested to variable degrees. In mobil belts, certain cycles are unrecognizable because they are united with related cycles, but they can be detected by statistical analysis of the planetary scale. Ronov and Khain (1970) and Seslavinskii (1987) have proved such cycles to be reliable using worldwide statistical data on lithologic complexes of the Hercynian–Alpine and Caledonian cycles, respectively. This is supported by much other data, including: statistics on unconformities (Pronin, 1973) and radiometric dating of granitoids (Zagruzina, 1987), analytical results on cycles in sedimentary deposits (by Afanas'ev) and on the distribution of glaucophane schists and ophiolites (Dobretsov, 1988), and data of seismic stratigraphy on sea level fluctuations (Haq *et al.*, 1987). For these reasons, there is no basis for rejecting the concept of global tectonic cycles.

This paper is based on our previous publications (Khain, 1971, 1977, 1979, 1984, 1985; Ronov *et al.*, 1984, 1989; Seslavinskii, 1987; Khain and Seslavinskii, 1990, 1991, 1992; Khain and Balukhovskii, 1993). They include references to books and papers published during the past 10 - 15 years and are used in this study together with our own data. There are thousands of works concerned with the problems under discussion, and we regret having no opportunity to refer even to all fundamental publications in this paper.

TECTONIC EVOLUTION OF THE EARTH DURING THE PALEOZOIC

Although our study is devoted to the Phanerozoic, we also need to briefly review the history of the Late Proterozoic in order to complete the desirable picture.

The existence of the supercontinent Pangea "I" can be referred to the period between 1800 - 1700 and 1350 Ma. Later in the Middle Riphean, it suffered rifting (mainly of continental, but in some areas of Red Sea type), accompanied by the formation of new oceanic crust. In the Late Riphean, especially after 900 - 850 Ma, this progressing process resulted in the appearance of many mobil belts with oceanic crust. Some of these belts related to northern continents of the Laurasian group, while others related to the future supercontinent Gondwana; the history of these two belt groups was different. In the latter regions, there was a distinct tendency towards the closing of these basins, followed by their Late Vendian transformation into fold systems, which served to join the Pangea "I" fragments into the new supercontinent Gondwana, matching nearly half of the next Pangea "II."

Events within the future Laurasia were different. Initially, the closing stage of the Late Riphean basins also occurred, followed by their transformation into fold systems. However, these processes were very quickly replaced by the destruction of continental crust and the opening of new, much larger basins with oceanic crust. By these means, paleooceans were formed: the Paleotethys separated the North European continents from Gondwana, the Iapetus (Paleoatlantic) formed between North America and Europe, and the Paleoasian ocean formed between the East European-Siberian and China-Korean (Sinian) continents. The existence of a separate Boreal ocean between Hyperborea (Arctida), Siberia, and North America is problematical. The formation of the paleooceans started in the Late Riphean time, was under progress during the Vendian–Cambrian, when the Gondwana was created, and experienced a mature stage in the Middle Ordovician. Thus, two continental segments of the Earth, Laurasia and Gondwana, went through Late Paleozoic evolution in different directions. In addition to this, there was the oceanic hemisphere of Panthalassa, the relic of which is represented by the present Pacific Ocean.

The mature stage of Gondwana evolution was during the Middle Paleozoic. During this time, the widespread marine environments resulted in the formation of a sedimentary cover on the plate, but tectonic and magmatic activity was at a minimum. The northern (relative to modern coordinates) margin of Gondwana, facing Paleotethys, included new crustal areas accreted in the Early Paleozoic. During the Middle Paleozoic, it was passive in contrast to the active Pacific margins (South America, Antarctica, and Australia).

At the end of the Carboniferous, Gondwana was in its late stage of existence. This was the stage when Gondwana and Laurasia were joining into the supercontinent Pangea "II." At the same time, this stage was preceding the Gondwana breakup, as is indicated by the rift system formation, which was most intensive in Southern Africa, India, and in some parts of Australia and Antarctica.

During the Riphean and Early Vendian, continental platforms of the Laurasian group suffered the avlakogene evolutionary stage and advanced to the plate stage earlier than Gondwana, before the Middle Vendian. Therefore, the first maximum of marine transgressions on the Laurasian continents was also earlier, occurring

during the Ordovician. In the Middle Devonian, platform cooling and subsidence was interrupted by tectonic and magmatic activity, which occurred most widely in Siberia and East Europe. It resulted in such events as: uplifts, rejuvenation of Riphean rifts, new rift formation, and the outburst of trap and alkaline basaltic magmatism. Before this time, many Laurasian continental margins were transformed into convergent boundaries, and North America and East Europe united to form one large continent, known as Euramerica or Laurussia.

After the Middle Devonian activity, coeval with the terminal collision in the Iapetus region, there was a new stage of subsidence of the East European, Siberian, and North American platforms during the Late Devonian and Carboniferous. Since the Permian, an opposite evolutionary trend caused the uplift and draining of the Laurasian platforms for the first time in North America. Before the end of Paleozoic, the platforms were surrounded nearly everywhere by rings of fold structures that were thrust over them.

Mobil belts framing the Pacific Ocean were the most active among the Paleozoic structures of this type. The initial Riphean–Vendian stage of their evolution was accompanied by the detachment of continental blocks of various sizes, and their subsequent transformation into microcontinents (Kolyma–Omolon, Sea of Okhotsk, central Kamchatka; also, Honshu and Kurosegawa in Japan, Arekipa in Andes, and others). Many island arcs were under development. Behind the arcs and microcontinents, there were marginal basins with crusts of transitional (Andes) or oceanic (Japan, northeastern Russia) types, the latter represented by ophiolites. Finally, the arcs and microcontinents collided with continental margins, resulting in continental growth and crust regeneration. Crustal accretion is most evident in southeastern China (Cataysia), eastern Australia (Tasman belt), and Antarctica (Rossides); this process was active during the initial and middle stages of the Paleozoic. The phases of accretion were not simultaneous, but occurred: at the end of Cambrian (Australia, Antarctica), at the end of Ordovician and in Silurian (Australia, Andes), in the Middle Devonian (Australia), at the Devonian–Carboniferous boundary (Antler orogeny of North American Cordillera and Central Andes), in the Middle Permian (Sikhote-Alin'), and at the Permian–Triassic boundary (Sonom orogeny of Cordillera, Australia).

Among intercontinental mobil belts, i.e., the paleooceans, the Paleoatlantic belt, or Iapetus, was the first to reach its final evolutionary stage. The origin of Iapetus is rather precisely dated by the Vendian–Cambrian boundary. However, it has now become more evident that in the Late Riphean–Vendian time, there was the more narrow and less deep Protoiapetus, or Celtic ocean located in the southeast. The Iapetus originated from a rift system of that age, which was striking parallel to the Protoiapetus on its northwestern side. At the beginning of Ordovician, the Iapetus had a maximum width that was estimated to be from 1000 to 3000 km, possibly even larger. The first phase of compression (the Oliver–Grampian–Finnmark phase) affected only the northwestern margin of the Iapetus and was related to an island arc or microcontinent collision with Laurasia. The island-arc evolutionary stage of the Iapetus occurred in the Ordovician. These island arcs were variable in their abundance and crustal type (sialic or simatic) along the belt's strike, and associated subduction zones had inclinations toward different sides.

The second, Middle to Late Ordovician phase of compression in the Iapetus region was most intensive in northern Appalachia and less designated in Newfoundland and the British Isles. This phase was caused by the collision of an island arc with the North American continent. A marginal basin was open behind this island arc during the Silurian, and then closed at the Silurian–Devonian boundary (Britain, Scandinavia). In Newfoundland and northern Appalachia, the last event occurred during the Middle-Late Devonian. From this time, the belt was at the orogenic evolutionary stage and then replaced by platform stages of the Late Devonian on the east and of the Early Carboniferous on the west.

The Urals–Okhotsk mobil belt, corresponding from a paleogeographic standpoint to the Paleoasian ocean, is wider than the Paleoatlantic belt and has a more complex structure and history. It includes large blocks of ancient continental crust of Early Precambrian age: the Kasakh, Tuva-Mongolian, Central Mongolian, and other microcontinents. The earliest stage of the belt's evolution lasted from the Middle through the first half of the Late Riphean (1350 - 850 Ma). It is adequately depicted in the Urals and Enisei Range. The next Baikalian stage, which terminated before the Cambrian time, is well represented in the Timan–Pechora, Polar Urals, and southern Barents Sea regions. It resulted in a certain crustal growth of cratons adjacent to the belt's periphery.

Beginning from the Altai–Sayan region, the eastern part of the belt is nearly free of Late Baikalides; the early evolutionary stage was quickly replaced here by the island arc stage of the Vendian–Early Cambrian, followed by the orogenic stage of the Late Cambrian. In the belt's western part, affected by the Late Baikalian orogeny, crustal destruction started in the Ordovician, and spreading processes were active until the Middle Devonian. Central Kazakhstan (except the Junggar–Balkhash region) and northern and partly median Tien Shan suffered intensive deformations and granitization at the end of Ordovician, while the Urals and southern Tien Shan went through the orogenic evolutionary stage only during the Late Paleozoic. The Altai–Sayan region, northwestern Transbaikalia, and northern Mongolia were strongly affected by the late Caledonian orogeny in the Devonian, during which even peripheral zones with earlier Baikalian and Salair deformations were touched by it. In the Devonian, the central zone of the Paleoasian ocean was transposed into the Ob'–Zaisan system, trending northward into West Siberia and to the southwest into southern and inner Mongolia, Dunbei, and Primorie. The eastern half of the belt is dis-

tinctly divisible into northwestern and southeastern zones. The first one suffered orogenesis in the Middle Carboniferous and the second, in the Middle Permian. The most eastern part of the belt, the Amur–Okhotsk system, existed as an open, deep basin – the gulf of the Pacific – until the end of the Middle Jurassic. In the entire latitudinal part of the belt from Tien Shan and central Kazakhstan to Transbaikalia, there were large volcano-plutonic zones of Andean type evolving along its northern periphery from the Devonian until the Early Cretaceous.

Judging from ophiolites of the Anti-Atlas and Arabian-Nubian shield, the intercontinental mobil belt of the Mediterranean, or the Prototethys, originated in the middle of Late Riphean (850 Ma) and terminated its first evolutionary stage before the Cambrian. The next Ordovician destruction left a wide zone of new continental crust, preserved to form an epi-Baikalian platform along the northern periphery of Gondwana. The platform includes Anatolia, Arabia, Iran, Afghanistan, northern India, and Sinoburmania. The Paleotethys, consisting of two basins separated by a chain of microcontinents, was formed in the European region before the Ordovician. The southern basin could be inherited from the Late Proterozoic Prototethys, as there are ophiolites of Vendian–Early Cambrian age. The axial chain of microcontinents was affected by the Devonian orogeny, which from here progressed northward and southward to reach the belt's margins in the Late Carboniferous.

TECTONIC EVOLUTION OF THE EARTH DURING THE MESOZOIC AND CENOZOIC

The Pangea "II" breakup is the main event in the Mesozoic–Cenozoic evolution of the Earth. Modern continents and the basins that separate them – the Atlantic, Indian, and Arctic oceans – were formed because of this breakup. The Late (or Middle) Triassic to Middle Jurassic continental rifting can be considered as the preliminary stage of the Pangea "II" breakup. Before its end, marine waters found their way into the rifts.

The first evolutionary stage of juvenile oceans started during the end of Middle Jurassic. It is recorded in a system of linear magnetic anomalies of Keathly set (M-29 - M-0) that were discovered in oceans and correspond to an interval that lasted through the Middle-Late Jurassic and the first half of Early Cretaceous. The stage should probably also encompass a quiet period without magnetic polarity inversions, corresponding to the Aptian–Santonian time. During this stage, the disposition of spreading centers in the Indian and Pacific oceans was distinct from later patterns, which were gradually progressing to the modern state.

The second stage in the history of young oceans started 90 - 80 Ma ago. During the initial period of this stage, many regions of Pangea "II" were still united, such as: South America and Antarctica; Greenland, Lomonosov Ridge, and Eurasia; and Madagascar and India. India was apart from Eurasia until nearly the end of the stage. Main modern patterns of the Earth's surface, including continental relief with present mountain ranges, appeared gradually after 45 Ma (geomagnetic anomalies 19 - 20) at the boundary between the second and third stages of development.

As was noted, the Pangea "II" breakup started as an immense system of continental rifts initiated during the Late Paleozoic. Since the terminal period of the Middle Jurassic, some of these rifts were transformed into intercontinental and then oceanic structures, while other elements of the system evolved without transformation until the Early Cretaceous. Coeval passive continental margins went through the rift and postrift stages with some time discordance in various marginal segments. In some places, their rift stage was characterized by basaltic volcanism, e.g., in northwestern Brazil it was active in the Late Jurassic time. Yet, magmatic activity was more widespread in the internal continental regions and left thick trap plateaus like those of the Rajmahal, Parana, Deccan, and other regions.

In the inner continental regions related to ancient cratons, some depressions, like syneclises of Paleozoic origin, were evolving with a gradual deceleration of subsidence rate. Young epi-Paleozoic platforms (West Siberian, Barents Sea, West European) were also subsiding from the Middle Jurassic until the Late Cretaceous–Early Paleogene after their rifting stage of Late Permian–Early Jurassic age.

During the Mesozoic and Cenozoic, the main mobil belts of the Earth were located along the margins of the Pacific and in the latitudinal belt of the Tethys. The continuous circum-Pacific belt did not exist during the Mesozoic, and its separate provinces were evolving independently. They can be divided into sectors according to their relations with margins of Eurasia (northeastern sector), Australia (southwestern sector), North America (northeastern sector), and South America and Antarctica (southeastern sector). In the Early Mesozoic, the Eurasian margin was discontinuous due to the South Anyui, Amur–Okhotsk, Qinling, and Laos–Vietnam basins with oceanic crust, which were open toward the Pacific Ocean. The basins separated continental blocks; however, before the Early Jurassic, the Qinling and Laos–Vietnam basins, with the Nadanhada–Ussuri Gulf, closed to form a continuous passive margin, traceable from the Bureya to the Indosinian massif.

Additional structural changes in the northwestern sector took place during the Late Jurassic–Early Cretaceous, when the Uyandina–Yasachii arc, and then the Omolon massif, collided with the Siberian margin. Coeval events included the closing of the South Anyui basin and the origin of the Khatyrka island arc. The marginal volcanic belt of the Okhotsk–Chukchi region originated in the Albian time.

The Cretaceous–Paleogene boundary marks the time of structural reorganizations in Koryakia, western Kamchatka, eastern Sakhalin, and central Hokkaido, where island arcs were coupled with the continent and

complex fold-nappe structures were created. The origin of the Central Kamchatka volcanic arc is also an event of this epoch. In the southern part of the sector, there was probably a collision between the West Philippine microcontinent and Asia. At this time, the paleo-Philippine volcanic arc was under formation, which was later followed by the opening of the West Philippine basin in the middle Eocene. In such a manner, the Asian margin of the Pacific was attaining features typical of the present West Pacific active margins with chains of volcanic arcs.

Many deep back-arc and inter-arc basins were opening during the Cenozoic: the above mentioned West Philippine and Sulawesi basins during the Eocene; the South China Sea during the middle Oligocene; the Japan, Sulu, South Okhotsk, and Komandor basins during the Miocene; and the West Mariana basin during the Pliocene. During the Miocene, these events of destruction were associated with opposite processes of accretion and collision. The latter resulted in obduction of oceanic complexes in the Olyutor zone of Koryakiya and eastern Kamchatka, as well as fold-thrust and slip deformations in eastern Sakhalin, central and eastern Hokkaido, the Shimanto zone of southwestern Japan, eastern Taiwan, and the Philippine Archipelago.

In the Triassic, the southwestern sector of the circum-Pacific belt served as the Gondwana active margin of the Andean type. All over this area the marginal volcano-plutonic zone, inherited from the Paleozoic, was evolving. The detachment of New Guinea was initiated in the Early-Middle Jurassic, and Australia was separated from southeastern Asia in the Late Jurassic. At the Jurassic–Cretaceous boundary, the New Caledonia–New Zealand segment was affected by Rangitata orogeny. In the middle of the Cretaceous, the Melanesian volcanic arc originated northeast of New Guinea. In the Eocene, it extended up to the Solomon Islands, and in the Oligocene – up to New Hebrides and Fiji.

An essential structural reorganization of the sector occurred in the terminal Eocene and Oligocene time. During this period, there were several events: the cessation of sea-floor spreading activity in the Tasman and Coral seas; ophiolite nappe obduction in Papua New Guinea, New Caledonia, and New Zealand; and the migration of a subduction zone from the Pacific to the Australian side of the Melanesian arc. A new epoch of tectonic activity near the Oligocene–Miocene boundary is depicted by: deformations in New Hebrides and Fiji, intrusive activity in Bismark and the Solomon archipelagos, the calc-alkaline magmatism in New Guinea, the origin of the Lau–Colville–Tonga–Kermadek volcanic arc, the Kaikura orogeny in New Zealand, and other events.

The present structural pattern of Melanesia developed in the terminal Miocene, Pliocene, and Quaternary. During this period, the New Guinea marginal sea, the Fiji Basin, and the Lau-Havre Trough were formed. Movements along large, sinistral strike-slip faults formed a zone of microplates; this is now bounded by a double chain of subduction zones with opposite inclinations and includes island arcs separated by marginal seas.

The Mesozoic–Cenozoic evolution of the southeastern sector of the circum-Pacific belt (Andes) was always related to an active continent–ocean boundary. For a short period of time (the Kimmeridgian–Valanginian), only the Patagonian Andes were similar in structure to active margins of the West Pacific type. The rest of the mobil system had the "liminar" (in Aubouin's terminology) arrangement. During the Triassic and Jurassic, the continental margin experienced a progressing subsidence accompanied by transgression. At this time, the margin was bounded on the oceanic side by an ensialic volcanic arc. A back-arc basin with oceanic crust appeared behind the southern most part of the Andes at the end of the Late Jurassic. On the south of the central Andes, a significant Kimmeridgian uplift transformed the island arc into a marginal volcano-plutonic belt, and submarine eruptions were replaced by acid terrestrial extrusions. Northward, similar processes were in action later, only at the beginning of the Senonian.

After a decrease, a new peak of endogenic activity occurred at the end of the Eocene. By this time, nearly all modern elements of the Andean structure had been manifested in the relief, and foredeeps were filling with molasses. At the Miocene–Pliocene boundary, nearly all intermontane troughs were drained, and the rising Andes experienced deformations, which also affected the Subandean troughs.

At the beginning of the Mesozoic, the general situation in the northeastern sector of the circum-Pacific belt, i.e., in the mobil system of the North American Cordillera, was defined by a deep basin that existed westward of the modern continental margin. This was a marginal sea that was separated from the Pacific ocean by island arcs, and conditions were similar to those of active margins of the West Pacific type. The sector went through a very complex evolution during the Mesozoic, when many terrains were joining to and slipping along the continental margin. These terrains represented fragments of various structural elements such as ensimatic and ensialic island arcs, oceanic crust of marginal seas, microcontinents, and intraplate oceanic rises. It is likely that since the end of the Middle Jurassic, the terrains coupled with the continent were displacing southward along a system of sinistral strike-slip faults. In the Late Cretaceous and Early Paleogene, they were displaced along dextral strike-slip faults backward to the north. The amplitude of the last displacement was higher than that of previous movements to the south. Thrust movements northward of the Brooks Range were active simultaneously, and the Rocky Mountains were also affected by thrust-fold deformations.

By the mid-Eocene time, the Cordilleran system had nearly attained its present pattern. However, at the beginning of the orogenic stage, it suffered significant transformations, probably related to the subduction of the East Pacific Rise under the continent. This stimulated several events such as: the formation of the San

Andrea's fault, the opening of the Gulf of California, the development of the Great Basin rift system, the initiation of the Rio Grande Rift, and basaltic magmatism activity in the Colorado Plateau. The widest part of the Cordilleran system experienced an extension that caused thinning of the crust and lithosphere, with the lower crust transposition under the Sierra Nevada and Colorado Plateau compensated for by a rising of the asthenosphere.

The history of the Mediterranean mobil belt is extremely complex due to the coexistence of various geodynamic environments in its different regions. A consequent detachment of microcontinents from the North Gondwana margin, followed by their coupling with southern Eurasia, defines the main evolutionary patterns of the belt. Accordingly, the Eurasian margin was growing due to the accretion of the Early and Late Kimmerian, followed by Alpine fold zones. Also, new basins with oceanic crust, located in a more southernly position each time, were opening and closing over time.

In the Triassic and at the beginning of the Jurassic, the western part of the Mediterranean belt was under a regime of the platform type. In the Carpathians-Balkan region, the regime was disturbed by deep zones with transitional and oceanic crust inherited from the Paleozoic. One of these zones extended from the Dobrogea through the Crimea Mountains, the Caucasus, the Transcaspian region, central Afghanistan, and northern Pamirs to the Kunlun Shan, Nanshan, and Qinling. On the northern side, it had a marginal volcano-plutonic belt and experienced compressive deformations after the Triassic.

Another chain of deep basins extended from the Slovakia–Hungary boundary toward the Wardar, Izmir–Ankara–Erzinjan, and Sevan–Akera zones, and further to the Iranian Karadagh, the Farahrud zone of Afghanistan, the Baggong–Nujian zone of Tibet, and central Myanmar. These basins became open mainly during the Jurassic time.

At the end of the Middle Jurassic, the western part of the belt split, initiating spreading and the formation of the Apennines and the Alpine mobil system. Eastward, in the Dinaro-Anatolian and Caucasus systems, compressive deformations of this age were accompanied by granitoid intrusions. At the Jurassic–Cretaceous boundary, these events affected a very broad region, extending from the eastern Alps to Myanmar, and destructed many deep basins. Simultaneously, new narrow zones with oceanic crust appeared in western regions such as the Alps, the Carpathians, and the southern Apennines.

During the Late Cretaceous, inner zones of the eastern Alps, Carpathians, Dinarides, Hellenides, and Balkanides experienced intensive thrust-nappe deformations with associated metamorphism and calc-alkaline magmatism. Before the end of the epoch, these zones came to their modern structural state, being thrust over outer zones that were still subsiding and accumulating flysch deposits.

The critical epoch in the Mediterranean belt evolution corresponds to the Eocene, during which the collision of the Indo-Australian plate and western microplates (Iberian, Apulian, and Bihor) with the Eurasian plate was initiated. Accordingly, the compressive deformations affected a wide region, extending from the Pyrenees and Maghreb to the Himalayas. At the beginning of the Miocene, the superimposed rifting resulted in the formation of a continental rift system in western Europe and the opening of the Algero-Provencal basin of western Mediterranean.

PHANEROZOIC GLOBAL RHYTHMS

Discussed structural changes of continents, oceans, and mobil belts resulted from many complex geological processes. Magmatic, metamorphic, and tectonic processes were primary among them. To analyze the evolutionary history of these processes, we compiled correlation tables. Two of them, containing data on mobil belts of the Mesozoic and Cenozoic, have been published earlier (Khain and Seslavinskii, 1990, Fig. 1; 1992, Fig. 1). The available information on 29 Paleozoic and 42 Mesozoic–Cenozoic mobil systems was analyzed at that time. Special symbols in the tables show age intervals of granitization, regional metamorphism, tectonic deformations, flysch and molasse accumulation, and volcanic activity of four major types. By the initial data processing, we compared the intensities of events in various regions and conventionally indicated them in the tables. All the results were used to approximate rates of intensity of main endogenic processes, which were responsible for magmatism, metamorphism, and tectonic activity variations (see next headings for discussion).

Data on the Phanerozoic magmatic and tectonic events in platform regions are schematicly presented in this paper (see below). This information was also used to evaluate global endogenic processes.

The geochronologic time scale after Harland *et al.* (1989) served as the basis for plotting our tables and histogram. By quantitative analysis, we used shorter (10 - 15 Ma) time intervals than those that are marked on the graphs. Some of these intervals correspond to an epoch or an age; however, longer epochs, e.g., the Early Carboniferous and Early Cretaceous, were divided into intervals equal to above values. The Early Triassic and Pliocene epochs, which lasted approximately 5 Ma each, represent the exceptions. The rate of the endogenic processes under discussion was estimated in conventional points for each interval obtained in such a manner (their total number for the Phanerozoic is 54). We then plotted corresponding diagrams.

OPHIOLITIC MAGMATISM

The term ophiolite means a sequence of serpentinized ultramafic rocks, gabbro, diabases, sheeted dykes, tholeiitic basalts, and siliceous sediments. Yet, full sequences are rare, and they usually occur dismem-

bered. Our tables (Khain and Seslavinskii, 1990, 1992) include only the most reliable data on tholeiitic basalts related to ophiolites. The symbol length in the tables corresponds to the time interval of rock formation, and a single symbol shows a very limited occurrence rate. The tholeiitic basalts are the best dated rocks of ophiolites, and we conventionally consider their ages as marking time boundaries for the entire ophiolitic sequence. To analyze the history of ophiolite development at a global rank, we summarized data from the tables and evaluated results according to a three-point scale by taking into account the number of structures bearing these rocks and their occurrence rate. For example, a single occurrence in the Early Cambrian of the Iberia–Atlas system and doubtful ophiolites (the Cambrian of the British zone) were evaluated by one point. Widespread occurrences during a certain part of the selected time interval deserved two points, or three points if they entirely covered this interval.

Of course, such gradations are very rough, and some values can be mistaken. However, we did not find an alternative approach for the unbiased analysis of very heterogeneous geological data, and we consider our approximations to be appropriate for global evaluations. The main results seem reliable, as there is a coincidence with previous compilations.

The results are shown in Fig. 1, where conventionally estimated values are plotted versus geological time. The variation curve revealed six epochs with a high occurrence rate of ophiolites: (1) the Early Cambrian, (2) the Early Ordovician and the first half of the Middle Ordovician, (3) the Early-Middle Devonian, (4) the Triassic, (5) the Late Jurassic and the first half of the Early Cretaceous, and (6) the Oligocene–Miocene.

The Early Cambrian epoch of ophiolite formation was initiated during the second half of the Late Riphean. The Late Riphean age (ca 850 Ma) is reliably estimated for ophiolites of the Mauritanides (western Africa), Anti-Atlas (El-Graara), and the Nubian–Arabian shield. Ophiolites, occurring along the boundary between the Bohemian massif and the Saxothuringian zone of the Central European system, and in the latter system, between the Münchberg massif and Gory Sowye, are all of the same or a somewhat younger age. Their western counterparts are represented by ophiolites of the Baja d'Adrienne in the far west of the Armorican massif and by ophiolites of the allochthonous sheets located in the Central Iberian zone of the

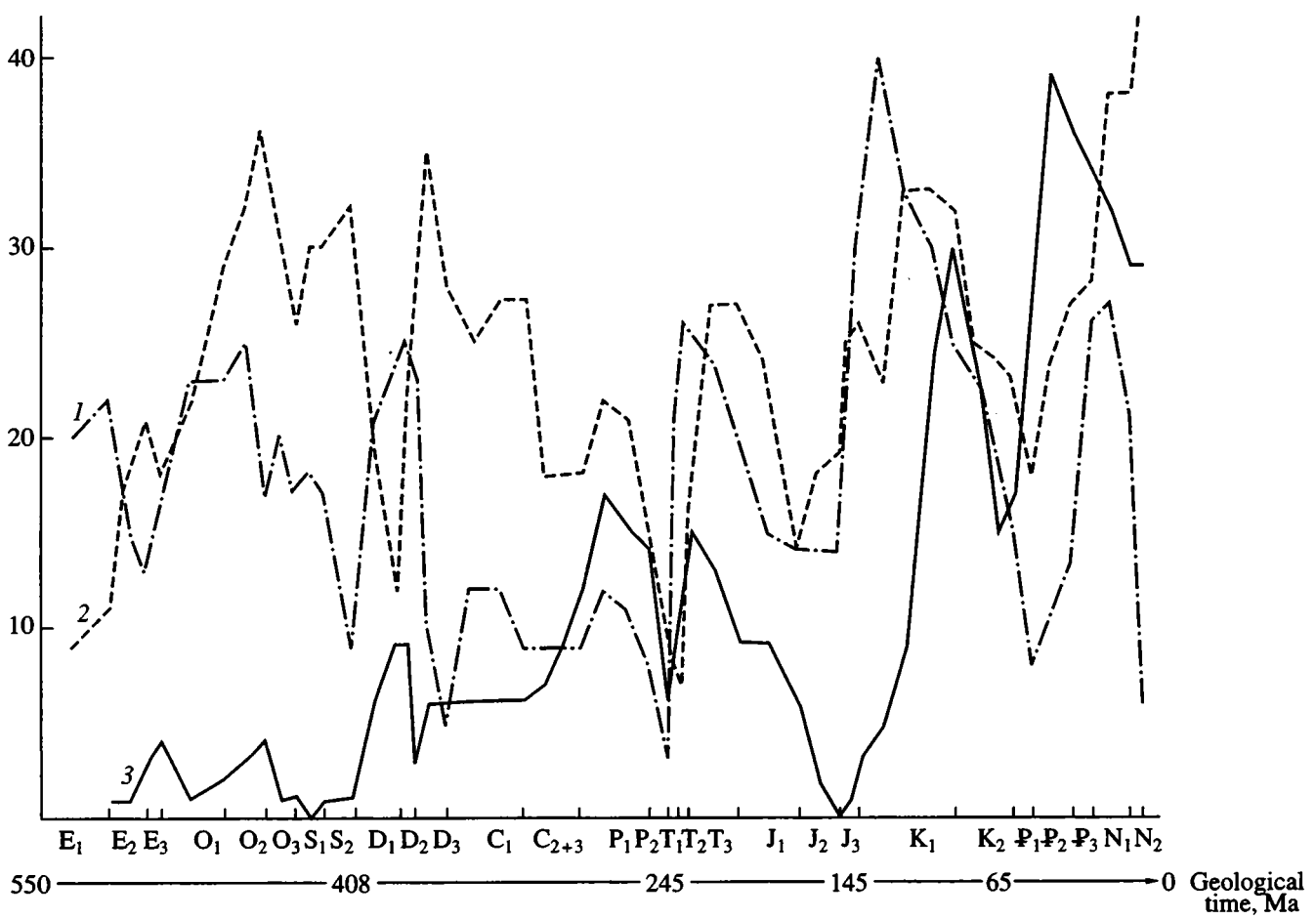

Fig. 1. (*1*) Secular variations in the abundance of ophiolites, (*2*) volcanic rocks of island arcs, and (*3*) marginal volcano-plutonic belts during the Phanerozoic (vertical scale – conventional points).

northwestern border of the Iberian Peninsula. Ophiolites of the age under consideration are present in the lower "diabas-phyllitoid formation" of Balkan Mountains, in the southern Carpathians (Corbu Group in the Iron Gates), and in Dinarides (the Wardar zone). Ophiolites of central Kazakhstan, the eastern Altai–Sayan region (West and East Sayan Mountains), and adjacent Mongolian regions also originated during this epoch. Pre-Ordovician ophiolites occur in the Cherskiy Range (northeastern Russia). Ophiolites in the western Tasmania and Lachlan system of eastern Australia could belong to the same epoch.

The next epoch, better constrained by datings, is represented by the Early-Middle Ordovician ophiolites, which are widespread inside the future Caledonian systems. These are ophiolites of the Appalachian, British, and Scandinavian Caledonides (the oceanic crust of the Iapetus), as well as rocks of central Kazakhstan and northern Tien Shan. Ophiolites of the earliest Ordovician age (496 ± 6 Ma) were recently recognized in the western Alps. In the western circum-Pacific belt, there are Penzhina ophiolites (northwestern Kamchatka) of this epoch. In the eastern circum-Pacific belt, coeval complexes occur in Oregon and eastern Sierra Nevada (American Cordillera), in the Argentinian pre-Cordillera, and probably in the Chilean Coastal Cordillera of the Andes.

The Hercynian systems reveal the third epoch of ophiolite development. Their Middle Devonian complexes are known in the Urals, in the Renohercynian (the Lizard Peninsula, Cornwall 375 Ma) and South Portuguese zones of the European Hercynides, on the north of the Innuitian system, in the southern Tien Shan and Pamirs, and in the Yunnan–Malayas system. In addition, this group includes Early Devonian ophiolites, occurring in the Mongol–Amur and Gobi–Hinggan systems (the Amur–Okhotsk zone), the Nippon–Sakhalin system, and the New England system of western Australia.

Ophiolites of the Triassic epoch are positively recognized in the northeastern part of the eastern circum-Pacific belt, namely in the Klamath Mountains and Sierra Nevada. The Cache-Creek ophiolitic rocks are presumably of a comparable age. In the northwestern segment of the circum-Pacific belt, the Triassic ophiolites occur in the Penzhina–Anadyr' and Amur–Okhotsk systems, Sakhalin and Hokkaido, and southwestern Japan (the Mikabu ophiolites). As for the Mediterranean belt, there are data on Late Triassic–Early Jurassic ophiolite occurrences in the Apennines, Dinarides, and Balkan Mountains, and probably also in the Lesser Caucasus and northern Anatolia. In the eastern part of the belt, this age can be suggested for ophiolites of central Afghanistan and Pamirs, and for those in the Early Kimmerian fold zones of China and eastern Indochina.

More confident information on widespread ophiolite development is available for the Late Jurassic–Early Cretaceous epoch. In the northwestern sector of the circum-Pacific belt, this generation includes ophiolites from the central Koryak Upland, the South Anyui system, the Sakhalin–Hokkaido region, and probably from the Shimanto zone of southwestern Japan and some regions of the Philippines. The eastern circum-Pacific regions, such as the Coastal Ranges of California, the Antilles–Caribbean system, Colombia, the Patagonian Andes, and New Georgia Island, also have ophiolites of this age. Ophiolites of the Papua New Guinea and northern Solomon Islands are their counterparts in the southwestern sector of the belt.

Ophiolites of the generation under discussion are widespread in the Mediterranean belt. They extend from its western segment (Beatic Mountains, Apennines, and Alps) through the Carpathians, Anatolia, and the Lesser Caucasus to the most eastern Indo-Burman regions.

It is likely that there was another epoch of ophiolite development at the beginning of the Late Cretaceous (80 - 75 Ma), but these complexes have a lower occurrence rate than rocks of the previous period, and both epochs seem closely related in time. The Late Cretaceous group includes some ophiolites from the northwestern (the Olyutor zone of Koryakia and western Kamchatka, Sakhalin, and Hokkaido) and southwestern (Papua New Guinea, Taiwan, and the Philippines) sectors of the circum-Pacific belt. In the central Mediterranean belt, ophiolites of this age occur from Cyprus and Syria on the west, to Iran, Oman, and Balujistan on the east. This epoch, or an even earlier one marks the formation of ophiolites in the Hindus–Tsangpo zone. The epoch came to an end at the Maestrichtian–Danian boundary.

The youngest Oligocene–Miocene epoch corresponds to the ophiolite formation widespread in the Melanesian sector of the circum-Pacific belt (Papua New Guinea, the Solomon Islands, and New Caledonia), the Philippines and eastern Taiwan, the eastern Kamchatka, as well as in California, the Antilles–Caribbean, and other regions. In the Mediterranean belt, these ophiolites occur mainly in its eastern part, beginning with Anatolia.

Ophiolites formed after the second half of the Oligocene are unknown in terrestrial occurrences, because the young oceanic crust suffered neither subduction nor obduction. However, ophiolites were under formation in the late Oligocene and during the entire Neogene, and are also developing at present. Spreading processes took place in the Mediterranean belt (the Algiers–Provencal and Tyrrhenian basins), the South Atlantic (Scotia Sea), the northeastern Indian Ocean (Andaman Sea), and many basins of the circum-Pacific belt (western Bering Sea, South China Sea, Sea of Japan, Philippine Sea, and others).

Thus, the epoch of the most intensive ophiolite formation corresponds to the Late Jurassic and Early Cretaceous. Peaks of the variation curve under discussion are separated by expressive minima that reveal not the cessation, but the decreasing rate of ophiolite formation. Fluctuations on the graph are large enough to be reliable. We believe that each epoch of intensive ophio-

lite formation corresponded to a global impulse of endogenic energy that resulted in the most significant structural reorganizations in the mobil belts of the Earth.

ISLAND-ARC VOLCANISM

Our tables also contain the available information on volcanic occurrences in the island-arc and continental marginal belts. Eruptions on island arcs were mainly of calc-alcaline composition. At the initial evolutionary stages of ensimatic arcs, volcanism was often tholeiitic in composition, while terminal stages produced subalkaline and alkaline extrusions. Some sequences differed, but we indicated all types by a single symbol.

In different mobil systems, periods of volcanic activity were variable, sometimes very short as in the Ossa–Morena zone of the Iberian Peninsula during the Middle Cambrian or the Afgano-Balujistan system during the Eocene. However, formation intervals of island-arc and marginal volcanic complexes were usually as long as tens of millions of years. Their most typical duration is equal to 30 - 40 Ma. This is well exemplified by the Jurassic island arcs of California and Alaska and structures of this type, which were evolving in the Sakhalin–Hokkaido system in the latest Cretaceous and Paleocene. The evolutionary period of the Okhotsk–Chukchi belt was also approximately 30 Ma.

However, some volcanic complexes have much longer periods of evolution. For example, the calc-alcaline volcanic eruptions in the Lachlan system of eastern Australia lasted from the end of the Middle Cambrian until the beginning of the Devonian (ca 120 Ma). The volcanic activity of the Sunda system, initiated during the Cretaceous, has been lasting for more than 90 Ma. The detailed data show that during such long periods, several generations of island arcs and marginal belts, which were active approximately 30 - 40 Ma each, replaced one another in an echelon manner.

Close relations between island-arc volcanism and subduction zones suggest that this type of volcanic activity is indicative of the same geodynamic environments in the geological past. Thus, a global analysis of secular variation in the distribution of corresponding volcanic complexes is very important for solving problems concerning the lithosphere tectonic evolution responsible for planetary tectonic phases of compression and extension.

Processing data on volcanism from our published tables, we used, as with ophiolites, the three-point evaluation scale. Local occurrences, such as one in the Eocene of the Carpatho-Balkan system or another in the Late Jurassic of the Amur–Okhotsk region, have been evaluated by one point. Volcanism of evolved marginal belts and island arcs deserved two points, when the activity period was shorter than a selected interval of the time scale, and three points, when the activity lasted for the entire interval. One point also corresponds to a doubtful volcanic occurrence. Despite the already noted defects in our approximations, we again consider the obtained global characteristics to be useful for attaining reliable results.

The curve of secular variations in the island-arc volcanism distribution revealed peaks of activity separated by epochs of lower occurrence rates (Fig. 1). The variation amplitudes are distinctly higher than fluctuations, which would result from our misinterpretations of the age and type of volcanic activity in 2 - 3 mobil systems. It is also remarkable that among Phanerozoic epochs, none are free of volcanic occurrences. Therefore, this activity was constantly present somewhere on the Earth, though with notable changes in intensity.

The first low peak correlates with the late Middle Cambrian (the Maya local stage). At this stage, island-arc volcanism was rather widespread in the Altai–North Mongolia regions, the Lachlan system of Australia, and some other locations.

The second peak marks the maximum activity period of the Paleozoic and corresponds to the Llandellian Stage of the Ordovician. During this time, island arc volcanic complexes originated within 13 of 29 mobil zones, especially in the North Atlantic and East Pacific belts, which have many volcanic arcs.

Later, until the Ashgillian, volcanic activity decreased and then rose again, which is marked by the third peak during the Late Silurian. In this epoch, volcanism was concentrated mainly in the Urals–Okhotsk belt (Urals, Kazakhstan–Tien Shan region), and also in the Appalachian system of the North Atlantic belt, the Lachlan system of the West Pacific belt, and other regions.

The next decline of island-arc volcanic activity lasted until the Emsian Stage of the Devonian and was deeper than in the Ashgillian. The fourth Frasnian peak of activity was accompanied by the spatial migration of volcanic occurrences. During this time they were concentrated in the central European, Caucasus–Pamirs, and Yunnan–Malayas regions of the Mediterranean belt. High activity also occurred in the Urals.

During the Late Paleozoic, island-arc volcanic activity was slowing, reaching a minimum at the beginning of the Mesozoic (the Early Triassic) when manifestations of volcanism were confined to the Sakhalin–Nippon and Cordillera systems. Insignificant rises are notable for the Namurian and at the beginning of the Permian period. The former is related to the island-arc volcanism of the Urals–Okhotsk belt, and the latter to the activity in the West Pacific belt.

The first distinct peak of Mesozoic volcanism is confined to the Late Triassic and the very beginning of the Early Jurassic. During this period, extensive eruptions of island-arc magmas occurred in the West and East Pacific belts and some areas of the Mediterranean systems (Yunnan–Malayas, Indo-Burman, and other regions). The next, somewhat lower peak appeared at the end of the Late Jurassic, mainly because of island-arc activity in the West Pacific belt (the Oloi, South Anyui, and other regions).

The following, more notable peak is close to the Late Jurassic one. Being weakly separated, both probably mark an integral stage of activity. Its maximum corresponds to the Aptian, when volcanism was especially intensive in the East Pacific belt (the Andes, Alaska, and Sierra Madre del Sur).

Later, the general occurrence rate of volcanic events under consideration was decreasing until the Paleocene, and rose quickly afterward to form the last peak. Let us remember that the data base for the Cenozoic includes information on the volcanic activity in modern transitional zones between continents and oceans, where all island arcs are now concentrated. Their eruptions and occurrences in many other systems (Sunda, Taiwan, the Philippines, New Zealand, and others) defined the highest intensity of the last peak, which is peculiar because we do not know whether it is at a maximum or not. Observing the extreme island arc activity at present, one can hardly predict what will occur in the future.

This activity is extreme in relation to the entire Phanerozoic. Events of a comparable rate were observed only in the Middle Ordovician. This should be kept in mind when one considers the evolution of the Earth crust, atmosphere, hydrosphere, and tectonic events from a standpoint of actualism.

Thus, periods of high and low activity of island arc volcanism were alternating. It is difficult to recognize distinct rhythms in this alternation. Intervals between described peaks lasted 40 and 60 Ma or even longer. Yet, certain peaks (the Maya, Late Silurian, Namurian, Late Triassic, Aptian and modern maxima) occurred with intervals of 100 - 110 Ma and seem to be periodic. Thus, the problem requires further study, although it is clear now that epochs of increasing volcanic activity mark global tectonic phases of subduction and dominant compression. The present epoch is most prominent in this respect.

It should be noted that these phases of compressive conditions are displaced in time, relative to phases of extension marked by epochs with high occurrence rates of ophiolites. The Middle Cambrian peak of volcanism (PV) followed after the Early Cambrian phase of ophiolite formation (OF), and so on: the Llandellian PV after Llanvirnian OF; the Late Silurian PV after Early Silurian (very weak) OF; the Frasnian VP after Eifelian OF; the Namurian PV after Visean OF; the Late Triassic PV after Middle Triassic OF; the Aptian PV after Neocomian OF; and finally, the present PV after Miocene OF. Only the minor Late Jurassic PV lacks a corresponding phase of ophiolite formation; in all other cases, the time shift is approximately 10 - 15 Ma.

The curve of secular variations in the volcanic activity of continental marginal belts revealed five peaks and a general trend of growth (Fig. 1). We do not know how reliable the last trend is, because difficulties in precisely dating terrestrial volcanic rocks are growing with their ages. Therefore, we do not exclude the possibility that the real degree of Paleozoic volcanic activity of continental marginal belts was higher than our approximation.

The first low peak of the Middle Devonian activity is related to volcanism of the north Appalachian and central Urals–Okhotsk belts. The second, of the Early Permian time, depicts events in Mediterranean regions and again in the Urals–Okhotsk belt. Next, the Late Triassic peak corresponds to activity in the Sikhote-Alin', the Andean system, and eastern Australia. Following this, the volcanism under discussion had nearly ceased before the Late Jurassic. In fact, the absence of any volcanic activity during the Middle Jurassic is enigmatic and lacks an adequate explanation.

After the Middle Jurassic, the calc-alcaline volcanism of continental marginal belts quickly progressed on the eastern part of the Asian continent and before the Late Cretaceous defined the fourth peak of activity of Albian age, when eruptions of planetary scale occurred in a vast area from the Okhotsk–Chukchi region to Kataysia in the south.

The fifth, also very sharp peak is confined to the Eocene, when corresponding volcanism spread over the Mediterranean belt (Pamirs–Hindukush, Transhimalayas, and Lesser Caucasus-Iranian systems), eastern Asia, the Andean system, and other regions. Afterward, the intensity of eruptions slightly decreased, yet its general level was still higher than in the Albian time.

Thus, the recent geological epoch has to be considered as unique in its intensity of volcanic processes on continents and continental margins. From the previous Aptian–Albian peak of volcanic activity, it is separated by an interval of approximately 100 Ma. If the revealed periods (100 - 110 Ma) of recurrent activity are reliable, we should expect even more tremendous volcanic processes on the Earth in the next several million years.

GRANITE FORMATION AND REGIONAL METAMORPHISM

Granitoid intrusions represent another form of magmatic activity, the data of which is also incorporated in our tables (Khain and Seslavinskii, 1990, 1992), though without rock classification by petrochemical types. This was unnecessary, because the study was aimed to reveal the most general regularity. Thus, special events such as plagiogranite or gabbro intrusions at the initial and mature evolutionary stages of mobil systems were not considered. We paid attention mainly to granite-leucogranite, leucogranite-alaskite, monzonite-granodiorite-syenite, granodiorite-granite, granite-migmatite, and gabbro-granite intrusive associations. Beyond the quiescent periods with zero value, epochs of granite formations were divided into three categories; intervals with minimum activity were designated by one point, and those with maximum occurrences by three points. The results of our approximations are given in the form of a histogram (Fig. 2).

In particular structures, periods of granite formation were variable, but two intervals, approximately 20 and

50 - 60 Ma, are the most characteristic. The first case is exemplified by the Middle Devonian phase of granitization in the Kazakhstan–Tien Shan region, or by that of the Early Cretaceous in the Oloi system. Longer periods were characteristic of events of the Carboniferous period in the central European regions, and of the Late Jurassic–Cretaceous interval in the Alaskan system. Yet, there were also very short events (e.g., the Famenian activity in the Cordillera system).

Periods of widespread activity of intrusive magmatism, which were recognized in five major mobil belts, did not coincide in time. These were the Silurian and Early Devonian in the North Atlantic belt, but the Carboniferous and Oligocene–Miocene in the Mediterranean belt. In the Urals–Okhotsk belt, there were two such periods, the Middle Devonian and the very prominent one of the Late Carboniferous–Early Permian time. Three periods – the Devonian, Late Permian–Early Triassic, and Cretaceous – were recognized in the West Pacific belt. The distinct Late Permian and Cretaceous periods were also revealed in the East Pacific belt, in addition to minor activity in the Late Silurian.

Despite these differences between the five evolutionary trends, the global synthesis revealed two distinct periods of very widespread granitoid intrusions – the Permian and Cretaceous – when processes of granite formation were active in all mobil belts. These were events of planetary rank.

The histogram also depicts other periods of more or less intensive granite formation. The first one corresponds to the Middle Cambrian, when granitoid intrusions were active mainly in the Kazakhstan–Tien Shan and Altai–North Mongolia regions, and in the Peru-Bolivian segment of the Andes.

The second period was longer, and lasted from the Late Ordovician until the end of the Carboniferous. It includes three active phases. First was the Ashgillian, when granite formation was most active in the Kazakhstan–Tien Shan, Kataysian, and Lachlan systems, but less significant in many other regions. The next phase of Late Silurian age also marks widespread occurrences of granitoids, with their maximum recognized in the Kazakhstan–Tien Shan region. The third phase corresponds to the Givetian and Frasnian of the Devonian. It is separated from the previous stage by a distinct decline in activity with a minor splash in the Emsian time. At the end of the Devonian, granite formation was most significant in the Appalachian and Lachlan systems.

The third period lasted through the Middle-Late Carboniferous and Permian and can be divided into the Late Carboniferous and Late Permian phases, the last of which was the most active of the entire Phanerozoic history. During the Late Carboniferous, the highest rate of granitization was in the West European regions. In the Late Permian, granites were under formation in 16 of 23 structural zones, which retained their mobility until the end of the Paleozoic. The process was most impressive in Central Asia, southern Tien Shan, and the Altai–North Mongolia region.

The fourth period of granite formation was very long. It started in the second half of the Late Jurassic

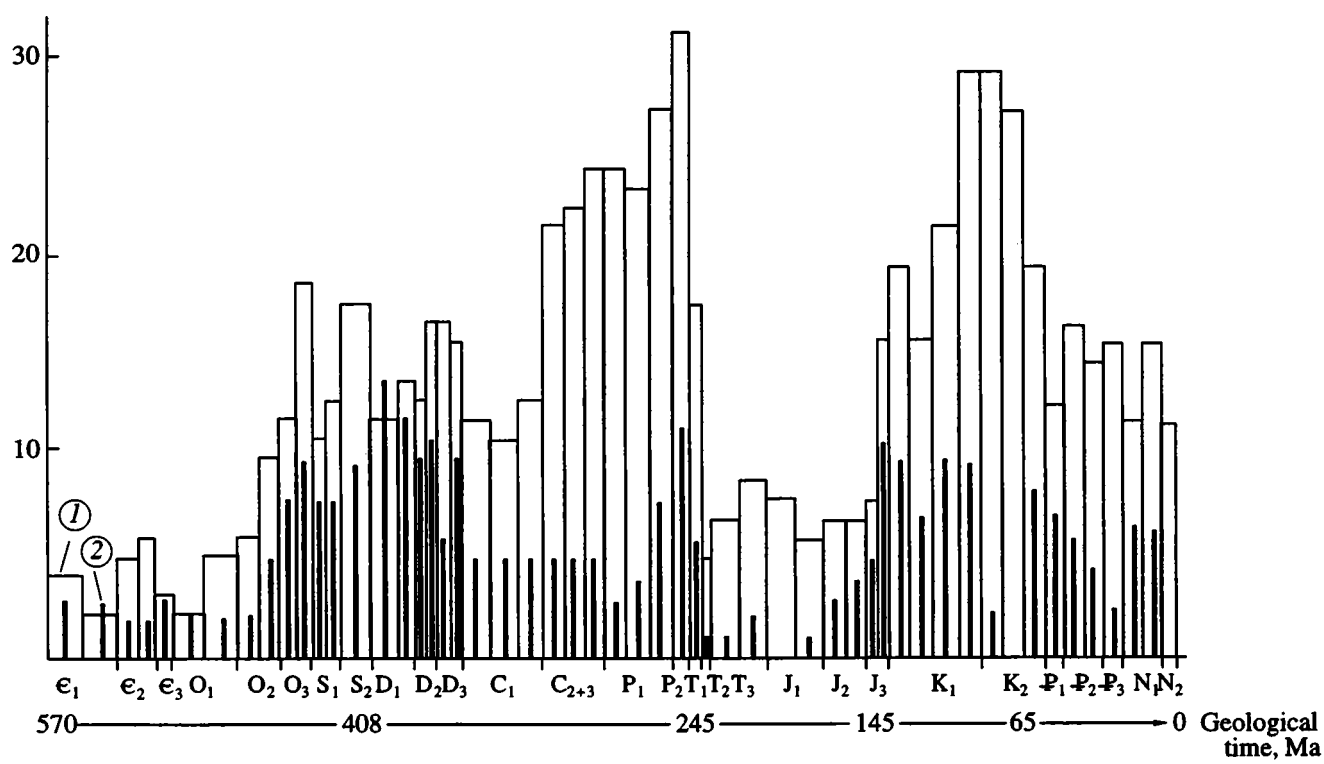

Fig. 2. Secular variations in the intensity of (*1*) granite formation and (*2*) regional metamorphism during the Phanerozoic (vertical scale – conventional points).

when granitization affected many systems (the Verkhoyansk region, Japan, Alaska, the Amur–Okhotsk and Pamirs–Hindukush systems, and others). A sharp decline in activity occurred at the Cretaceous–Paleogene boundary, when granitization processes ceased in the Asia Minor region and the northwestern sector of the West Pacific belt. The fourth period is approximately 80 Ma long and probably includes two superimposed shorter periods. A possible boundary between them is marked by a minor activity decrease in the Hauterivian and Barremian (120 - 130 Ma); the first part includes the Late Jurassic and Neocomian, while the second one lasted from the Aptian until the end of the Cretaceous. During the activity maximum at the beginning of the Late Cretaceous, granitoids intruded 16 of 42 mobil systems. This was the culmination period of collision in structural zones framing the Pacific Ocean.

A long, interesting period of rather insignificant granite formation lasted from the Middle Triassic until the Late Jurassic (ca 90 Ma). Granites of Late Triassic age are known only in two or three regions. Later, granitoids of the Qinling and Malayas systems slightly elevated this level of granitization. Being combined with data on the very low occurrence rate of ophiolites and island-arc volcanic complexes of Middle Jurassic age, this fact suggests a global decline of tectonic activity in this period.

The Cenozoic history of granite formation is difficult to understand, because the level of erosion in younger mobil systems is insufficient. Still, the exceptionally widespread occurrences of recent island-arc volcanism provide the grounds for assuming an intensification of granite forming processes during the Neogene.

The Cretaceous period of extensive granitization is separated from the preceding Permian epoch by an interval of 120 Ma. Similar intervals also separated the corresponding periods in the Paleozoic, and if this regularity is valid, our planet is now approaching the next stage.

Shorter phases of granitization followed one another after intervals of 40 - 45 Ma. On the histogram, they correspond to: the Thommotian, Middle Cambrian, Arenigian, Late Ordovician, Late Silurian, Late Devonian, terminal Early Carboniferous, Carboniferous–Permian boundary, terminal Permian, terminal Late Triassic, Middle Jurassic (a very weak phase), Neocomian, Albian, initial Eocene, and terminal Miocene. The periods between them (40 Ma) correspond to rhythms of island arc volcanic activity and global tectonic cycles. The Middle Paleozoic intensification of granitization terminated the Caledonian cycle; the Permian one was final in the Hercynian cycle; the maximum at the Jurassic–Cretaceous boundary depicts the end of the Kimmerian cycle; and the recent epoch probably indicates termination of the Alpine cycle.

Events of regional metamorphism of the epidote-amphibolite and higher facies occurred under similar paleotectonic environments and in the same regions, which were affected by magmatic activity. We evaluated only significant events that were active over large regions and associated with the emplacement of granite-granodiorite plutons. Periods of metamorphism were of various duration, from several to 20 - 30 Ma, such as in Japan at the Jurassic–Cretaceous boundary. Periods of metamorphism were usually shorter than those of granitization. In particular regions, metamorphic, magmatic, and tectonic events were interrelated in a complex manner, though metamorphism often occurred before granitization.

The generalized evolution of regional metamorphism, as is shown in Fig. 2, represents only a conventional approximation. The detailed quantitative analysis of reliably dated metamorphic complexes is an important goal of further studies. In this preliminary work, we estimated only the well known facts; when metamorphic processes lasted throughout the entire selected interval of time, we evaluated their intensity by two points, and only by one point, if their duration was shorter.

As was expected, the resulting histogram revealed a general coincidence between periods of metamorphism and granitization. The first Late Ordovician–Devonian period corresponds to the time of termination of the Caledonian tectonic cycle of Earth evolution. During its culmination in the Early Devonian phase, regional metamorphism was in action all over the North Atlantic belt, as well as in the Qilianshan, Urals, Kazakhstan–Tien Shan, Lachlan, and Kataysian systems, and in the Equador–Colombia segment of the Andes. According to paleomagnetic data, this was a time of an accelerated convergence rate of the Laurasian continental blocks.

The second period at the end of the Permian was less significant despite the extremely intensive granitization occurring during this time. In this period, metamorphism spread over the East Pacific belt (Cordillera, Andes) and over the New England, Yunnan–Malayas, and other systems.

After the Triassic decline and low activity in the Early-Middle Jurassic, there was first Mesozoic culmination of metamorphic events in the Late Jurassic–Neocomian (the Lesser Caucasus, Iran, Penzhina–Anadyr', New Zealand, Cordillera, Kataysia, and other systems). Next, culmination soon followed in the Aptian–Albian time, when metamorphism again was active in the Lesser Caucasus, Cordillera, and Kataysia, as well as in the Sunda, Taiwan–Philippines, and Equador–Colombia systems.

Another period of activity was in the terminal Cretaceous and Paleogene, when metamorphic processes affected the Nippon, Sakhalin–Hokkaido, Alaskan, and Antilles–Caribbean systems. The last culmination of the Miocene mainly influenced the Mediterranean belt. It is likely that metamorphic processes were and are active in present mobil systems of the West Pacific belt. Here, their culmination could be even more significant than the previous one; yet, unfortunately, recent metamorphic complexes are still unexposed by erosion.

TECTONIC ACTIVITY ON PLATFORMS

Ideas concerning very weak tectonic activity within platform regions need to be revised in relation to recent data. In this section, we will analyze results of tectonic processes that affected all platforms during the Phanerozoic in order to reveal the general evolutionary trend of the endogenic regime of these planetary structures.

We processed corresponding data on ancient structures with the Precambrian basement (Siberian, East European, China-Korean, Yangtse, Indian, Australian, African, North and South American, and Antarctic platforms) and their younger epi-Caledonian and epi-Baikalian zones (Dunbei, West Siberian, West European, Turan, East Australian, Scithian, central Kazakhstan, and Patagonian systems). The resulting correlation table (Fig. 3) shows the estimated time intervals of such processes as: riftogenesis, trap formation, alkali-basalt and kimberlite magmatism, granitization (anorogenic A-type granites), and alkaline-ultramafic intrusive activity.

Generalized secular variations were analyzed only for riftogenesis, trap formation, and alkaline magmatism, since other events, such as anorogenic granite emplacement, have a very limited occurrence rate. The evaluation procedure was similar to that applied for mobil belts, but simplified because available scarce data does not allow for the analysis of events in particular structures. Therefore, any event confidently known for a selected interval of time was evaluated by one point, and then their sums were used to plot variation curves (Fig. 4). Only two doubtful rifting events, suggested for the Early and Middle Paleozoic history of the African and Australian platforms, were marked by the lower value of 0.5. Thus, all these approximations are conventional.

Rift zones periodically originated on platforms during the entire Phanerozoic as a result of tectonic environments related to crustal extension. In particular structures, intervals of riftogenesis were highly variable, sometimes as long as 80 - 85 Ma, as in the Rocky Mountains and the South Oklahoma region of the North American platform. Longer periods of rift formation, shown in Fig. 3, correspond to all rifting events in various regions of the West European and African platforms. In Europe, these events formed the latitudinal rift system of the North Sea region in the Late Permian–Early Jurassic time and the Danish-Polish avlakogene of the Triassic–Early Jurassic period. In Africa, there were formed: the Central African rift system of the Late Jurassic–Early Cretaceous period, the Early Cretaceous West African rifts, the Bay of Syrte rift of Late Cretaceous–Paleogene origin, and the Paleogene–Neogene rift belt of the Afro-Arabian region. A minimum duration of complex processes responsible for riftogenesis is approximately 20 Ma; this is exemplified by many structures (the Middle Jurassic Carnarvon Rift in Australia, and others), but the most typical value is approximately 50 Ma.

As was already noted (Milanovskii, 1983), Early Paleozoic riftogenesis was unusual, and its manifestations were limited (see Fig. 4). We recognized corresponding events in North Africa (the Ougarta avlakogene), Arabia (the Palmyra avlakogene), and Australia (the Amadeus avlakogene, and also Perth, Carnarvon, and other troughs on the west of the platform). In addition, marginal riftogenesis affected the frontal Cordillera zone in North America during the Early-Middle Cambrian, and the Silurian Oslo graben was formed in northern Europe at the beginning of the Middle Paleozoic.

The first notable intensification of Phanerozoic rift-forming processes was during the Middle Devonian, when the Vilyui (Siberia), Dnieper–Donetz, Pripyat', Vyatka, Don–Medveditsa, and Pechora rifts were under formation in the East European platform. In addition, the Ougarta rift was formed in Africa, the Nandan and Qinzhou rifts appeared in the Yangtse platform, and several troughs originated in central Kazakhstan. The evolution of many rifts terminated before the end of the Devonian; some of them (Donbas) were filled with sediments and underwent compressive deformations during the Carboniferous. Before the mid-Carboniferous time, rift-forming processes ceased everywhere.

The next impulse of riftogenic activity started during the Late Carboniferous and culminated in the Artinskian and Kungurian of the Permian. This was a time of the active evolution of the following structures: Oslo Graben, Narmada–Son–Damodar graben system (India), Fitzroy Graben (Australia), Zambezi–Ruhuhu graben system (Africa), rift systems of the Rocky Mountains and Oklahoma (North America), Amu Darya rift (Kazakhstan), and many other grabens. These structures were concentrated mainly in the eastern parts of the Gondwana supercontinent.

The third, most prominent peak of activity is confined to the Late Jurassic–Neocomian time. Many of the platforms under consideration experienced extension and riftogenesis during this period. Among them there were: the China–Korean platform affected by the Early Tien Shan rifting on the north and northwest; the Indian platform with rifts originated in its southeastern part; the African platform with central and western rift systems; the South American platform with Salado, San Jorge, and Colorado rifts; the West European platform with rifts of northern Germany and the North Sea; and the Scithian platform with the Sivash rift system. Like the previous one, this culmination was again related to the Gondwana, but now both to its eastern and western regions.

The evolution of present rift zones generated in the Oligocene was under progress in the Neogene, when riftogenesis was intensified on platforms of both the Gondwanan (rifts of eastern Africa, Arabia, and Antarctica) and Laurasian (rift systems on the east of the China platform and on the west of the North American platform) groups. We suggest that these processes are still active at present and are still below a possible peak of activity.

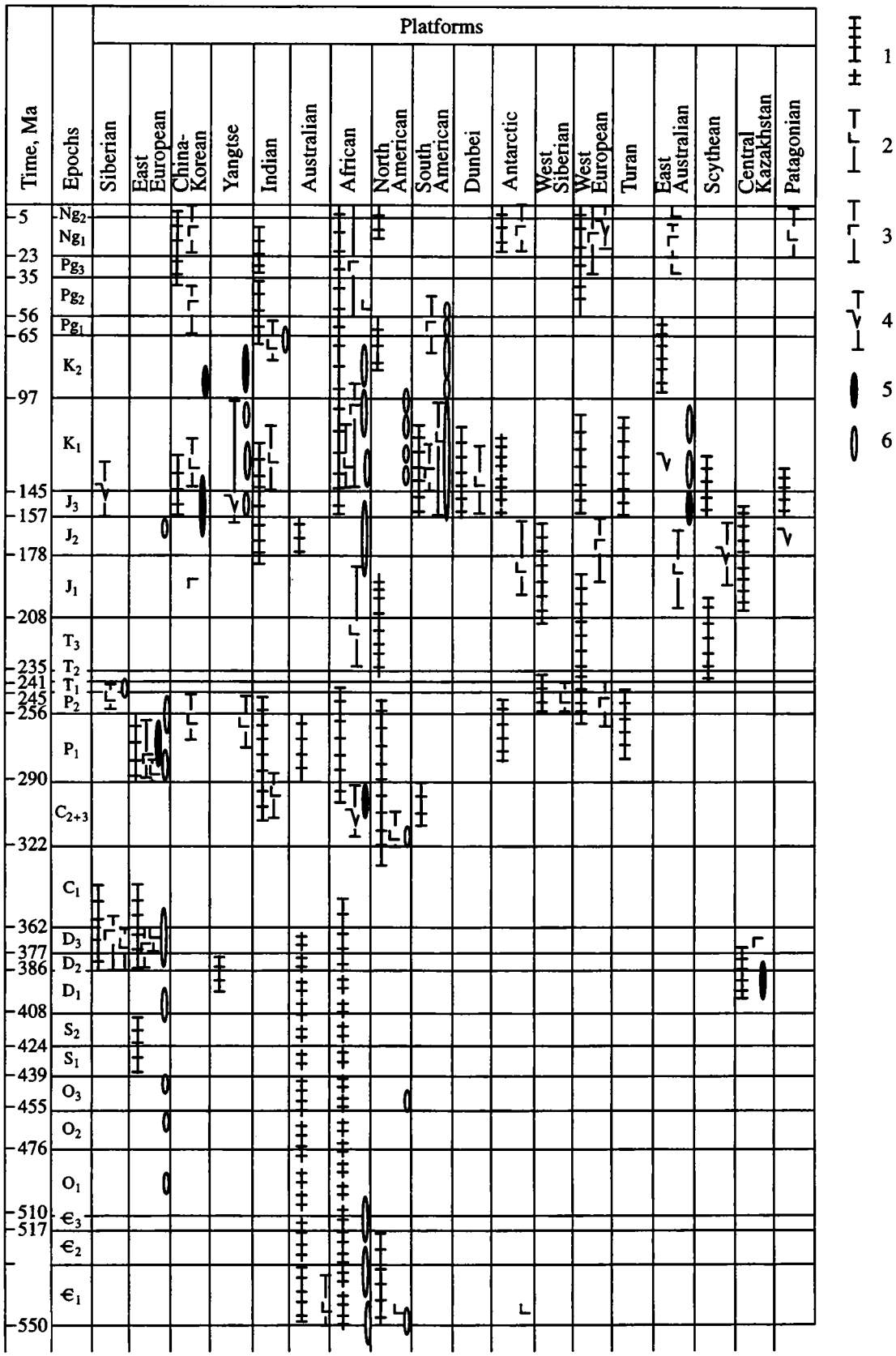

Fig. 3. Correlation scheme of Phanerozoic processes of rifting and magmatism on continental platforms of the Earth.
1 – riftogenesis (dash lines for suggested events); 2 – trap formation; 3 – eruptions of alkaline basalts; 4 – eruptions of acid and intermediate lavas of alkaline and calc-alkaline series; 5 – intrusions of A-type anorogenic granitoids; 6 – intrusions of alkaline and ultramafic magmas.

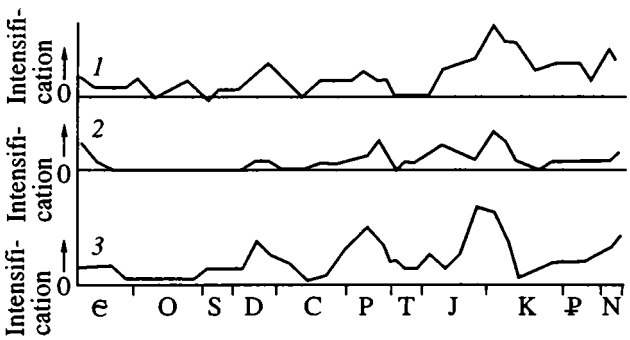

Fig. 4. Secular variations in the intensity of rifting and magmatic processes in all platform areas of the Earth during the Phanerozoic: (*1*) riftogenesis; (*2*) trap magmatism; (*3*) alkaline magmatism (vertical scale – conventional points).

Our analysis revealed two essential events of Phanerozoic trap formation: the first related to the Late Permian–Early Triassic period, and the second occurred in the Early Cretaceous time (Fig. 4). The former depicts the trap formation in Siberia and the Tarim block, as well as on the Sikan–Yunnan uplift of the Yangtse platform and on the north of the North American platform. The latter corresponds to the Neocomian trap magmatism on the north of the China-Korean platform, in the Namibia region of the African platform, and in the Parana and Maranao basins of the South American platform. Both maxima of events correspond in time to two culminations of rift-forming processes.

Trap formation was not characteristic of the larger period of the Paleozoic. The known occurrences are represented by the Lower Cambrian rocks of the Antrim Plateau in Australia and the Middle-Upper Devonian traps of the Vilyui (Siberia) and Dnieper–Donetz (East European platform) avlakogenes. There were also manifestations in the Late Paleozoic (Fig. 3). In contrast, the Mesozoic era was more liberal in this respect, which corresponds to higher tectonic activity on platforms stimulated by the Gondwana breakup.

In particular regions, intervals of trap formation were dissimilar. Recent data on large and well studied trap provinces show that, despite large volumes and areas of activity, periods of lava eruptions and intrusions of dykes and sills were very short. The period of trap magmatism in the Deccan Plateau (India) is almost precisely confined to the Cretaceous–Paleogene boundary (Gavey, Courtillot, 1989). In the Tungus province of Siberia, available trap ages correspond to the Permian–Triassic boundary age, with deviations of approximately 4 Ma (Campbell *et al.*, 1992). According to radiometric dating, rocks of the Parana province originated at the Jurassic–Cretaceous boundary (Renne *et al.*, 1992). Unfortunately, we were late in including this data in our correlation table, which shows larger periods of trap magmatism estimated earlier for the Deccan and Parana provinces.

We also analyzed secular variations of alkaline magmatic activity on platforms in combination with the occurrence rates of ultramafic complexes and calc-alkaline volcanic rocks. Time dependent variations of such a combined magmatic activity on platforms are shown in Fig. 4.

The plotted variation curve revealed a weak background intensity of the magmatic activity under examination and three periods when it was somewhat higher than usual. The first period of the Middle-Late Devonian time is related to magmatic events in the Vilyui avlakogene in Siberia and the Dnieper–Donetz and Pripyat' avlakogenes of the East European platform. Kimberlites of the Arkhangel'sk region were also formed during this period. During the second Permian period, the magmatic activity was concentrated on the north of the East European platform, in the Oslo graben region, and left alkaline intrusions and basaltic lavas there.

The third, more active period was in the Early Cretaceous. It was preceded by the Middle-Late Jurassic intensification of magmatic activity in Africa (alkaline massifs of the Jos Plateau, Nigeria, and western Africa), North America (alkaline intrusions on the southeast), the Lower Yangtse province (alkaline lavas and intrusions), and some other areas. In the South American and Yangtse provinces, the activity of alkaline magmatism lasted during the Early Cretaceous, when it also spread over the Arabian-Nubian arch, North American regions between the Ottawa and St. Lawrence grabens, and some regions of the young East Australian platform.

The next significant intensification of magmatic activity originated in the Neogene. It resulted in widespread eruptions of alkaline basalts in the Afro-Arabian rift belt, and the rift systems of eastern China, the Lower Rhine graben of Western Europe, Eastern Australia, and some regions of Antarctica. We suggest that there also were Neogene alkaline intrusions that are still unexposed by erosion. In that case, the estimated range of platform magmatic activity would be greater.

Hence, against a background of very stable tectonic conditions long lasting on platforms, there were also periods when tectonic and magmatic activity was distinctly intensified, presumably by global endogenic perturbations. There were three periods of this kind – the Middle-Late Devonian, the Permian, and the Late Jurassic–Early Cretaceous – all manifested on the surface by processes of extension, riftogenesis, trap formation, and alkaline magmatism activity. Another, more weak but still distinct global manifestation of these processes is confined to the recent epoch. They are still under progress, and their future trends can hardly be evaluated.

The evident correlation between the three variation curves shown in Fig. 4 suggests that the considered tectonic and magmatic events in platform regions were interdependent. It is remarkable that peaks of riftogenic activity were 15 - 20 Ma ahead of the impulses of the two other processes. In general, the recognized peaks of

events are separated by intervals of approximately 120 Ma, and this frequency is presumably dependent on the pulsation of endogenic activity of the Earth. Spatial distribution of individual activated regions can be dependent on both the mantle convection patterns and the stress field in the crust of a platform.

TECTONIC DEFORMATIONS IN MOBIL BELTS

Analysis of the paleotectonic history of continents, mobil belts, and oceans supports the well-known idea of the "intermittently continuous" evolution of tectonic processes. Tectonic movements permanently acted on the Earth; however, there were particular episodes of intensive tectonic activity that resulted in significant qualitative changes in structural patterns of the tectonosphere. Lesser peaks of tectonic activity in limited regions, which lasted less than 1 - 2 Ma, we consider as tectonic phases. Longer periods of approximately 15 - 20 Ma, which include several inner phases and were somewhat synchronous at the planetary rank, correspond to tectonic epochs. A group of tectonic epochs represents a tectonic era, which usually lasted approximately 150 Ma, and is equivalent to the Bertrand's tectonic cycle (the Caledonian, Hercynian, Kimmerian, and Alpine).

Below, we present a brief review of the main tectonic deformations of the Phanerozoic interval in order to recognize principal tectonic epochs and eras in the Earth's history, i.e., to reveal the global periodicity of tectonic evolution. In this review, we used the correlation tables of tectonic, magmatic, and metamorphic events described in previous sections.

In addition to the other information, these tables show age intervals and intensities of tectonic deformations in 29 Paleozoic and 42 Mesozoic–Cenozoic mobil systems, forming structural patterns of Phanerozoic mobil belts. Processing the available information, we estimated the number of systems affected by deformations in a particular epoch and the intensity of these deformations by using a four-point evaluation scale. Weak deformations that lasted shorter than a selected time interval were evaluated by one point; an example of this is the Early Cambrian folding in the Anti-Atlas. If such deformations lasted throughout the entire selected interval, as in the Alaskan system during the Paleocene, they deserved two points. Three points indicated deformations of moderate intensity during the entire interval under consideration (e.g., the Miocene folding of the Taiwan–Philippines system). A strong (main) folding phase of similar duration, like that of the Early Permian in the northern Andes, deserved the highest value of four points. A shorter duration of a moderate or strong folding phase reduced its value by one point. Of course, this evaluation system is primitive, and some values can be mistaken. Yet unfortunately, it is impossible so far to obtain more confident quantitative data, and we hope that our approximations reveal at least the most general regularities.

First, the resulting histogram shows a distinct pulsation of tectonic processes (Fig. 5), when peaks of their activity are separated by depressive periods. Against this pulsating background, there are periods of concentrated epochs of tectogenesis, which correspond to the intermediate and terminal stages of the Caledonian, Hercynian, Kimmerian, and Alpine tectonic eras. The first, earliest period lasted from the terminal Ordovician until the Carboniferous; the second, from the mid-Late Carboniferous through the Permian; the third, from the terminal Jurassic through the Early Cretaceous; and the fourth, from the terminal Eocene through the Oligocene and Neogene. Let us consider them in more detail.

At the very beginning of the Paleozoic, indications of some tectonic activity are notable in the mid-Cambrian time, when the Maya Stage marked the culmination of the Salair epoch of tectogenesis. In this epoch, deformations were most intensive in the Kazakhstan–Tien Shan and Altai-North Mongolia regions. It is remarkable that peaks of granitization and island-arc volcanic activity (Fig. 1) are also confined to the end of the Middle Cambrian. Therefore, there is no doubt that the Salair tectogenesis was of a global rank and marks a compressive stage in the Earth's history. The Grampian epoch of the Tremadocian caused deformations in the Britain, Scandinavia, and Lachlan systems of Australia and was of a comparable rank.

The first Phanerozoic period of strong intensification of endogenic activity originated after some decrease in the Arenigian and lasted until the end of the Devonian, encompassing several epochs of tectogenesis. The first of these is the Taconian orogeny, with culmination in the Ashgillian. Strong deformations of this time affected the Appalachian, British-Scandinavian, and Lachlan systems, along with the Kazakhstan–Tien Shan region, the Chile-Argentinian segment of Andes, and other zones. In addition, it was an epoch of intensive metamorphism and granite formation.

The next Ardennian (Scandean) epoch of tectogenesis is confined to the Late Silurian. It is very distinct as the terminal epoch in the Caledonian era of tectonic evolution. During this time, folding and other deformations spread over all mobil belts, being most energetic in the British-Scandinavian and Lachlan systems and in the Kazakhstan–Tien Shan region. The Late Silurian is also peculiar because of associated peaks of granite formation and island-arc volcanism, which are coupled with a sharp minimum in ophiolite formation history (Figs. 1 and 2). Also, this was a time of global regression and high sedimentation rates (Seslavinskii, 1987). The combination of above characteristics marks the Ardennian epoch as a very significant one in the general tectonic evolution of the Earth; it was definitely a planetary epoch of compression.

As compared with the Late Silurian, the beginning of the Early Devonian corresponds to a quieter tectonic period. Tectonic processes ceased everywhere in the Urals–Okhotsk belt. In many other regions, granitization was limited, but ophiolite formation was rapidly progressing. As an exception, we have to note the regional folding in the Central European system, which

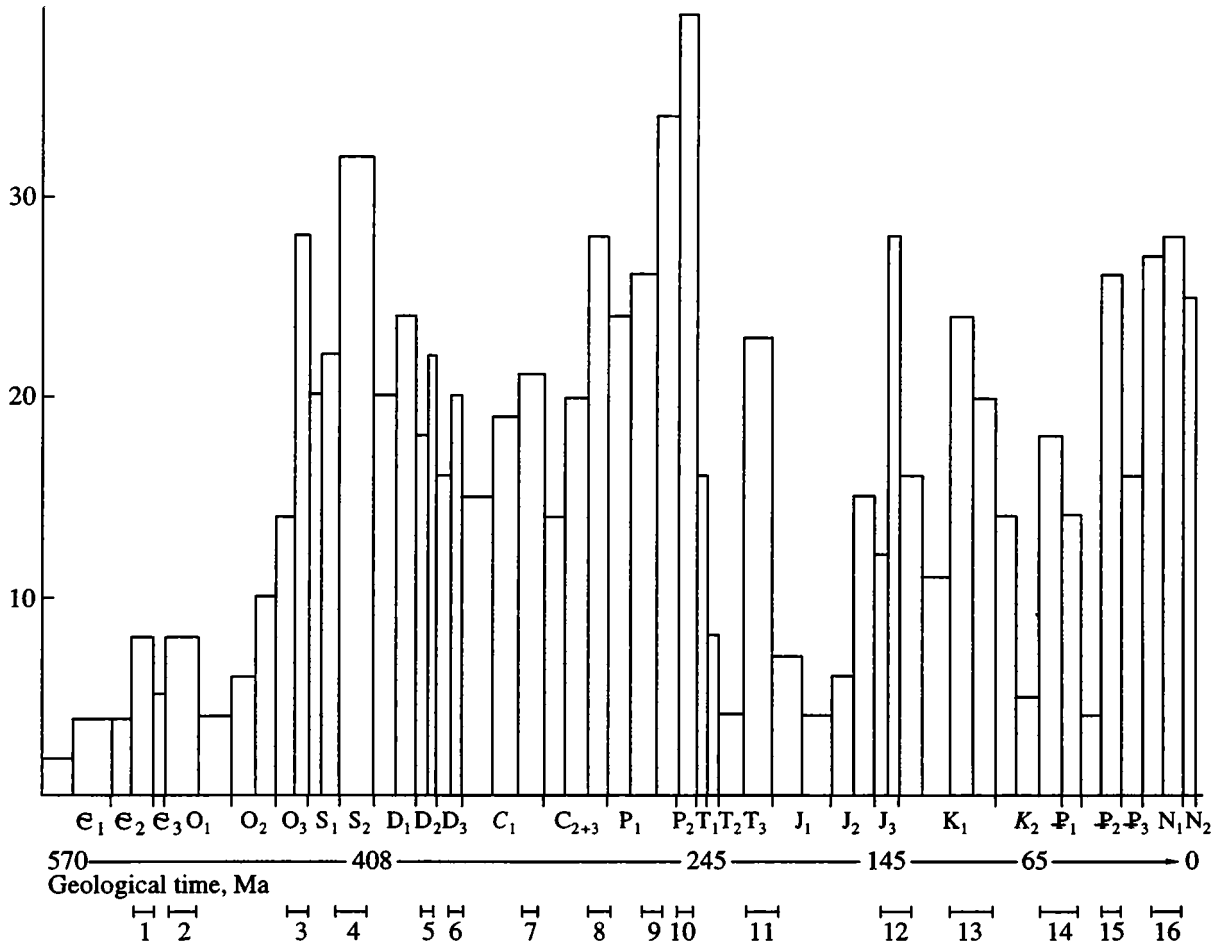

Fig. 5. Secular variations in the intensity of Phanerozoic tectonic deformations (vertical scale – conventional points).
Numbers underneath mark epochs of tectogenesis: 1 – Salair; 2 – Grampian; 3 – Taconian; 4 – Ardennian; 5 – Acadian; 6 – Bretonian; 7 – Sudetic; 8 – Asturian; 9 – Saalic; 10 – Pfalzian; 11 – Early Kimmerian; 12 – Late Kimmerian; 13 – Austrian; 14 – Laramide; 15 – Pyrenean; 16 – Attic.

affected the Caledonian volcano-sedimentary complexes before the end of the Gedinnian. The Emsian of the Early Devonian is controversial in regard to its paleotectonic characteristics. The intensified folding and granitization suggest the Emsian to be another epoch of tectogenesis. Yet, it was also a time of very depressed island-arc volcanism and elevated rates of ophiolite formation. Thus far, this combination remains enigmatic.

The next Acadian epoch of tectogenesis, which culminated in the Givetian age, is more distinct. During this time, deformations affected Appalachia, Scandinavia, central Europe, eastern Australia, and other regions. For the period under consideration, the last epoch of tectogenesis can be recognized as corresponding to the Famennian of the Late Devonian. It is adjacent in time to the traditional Bretonian orogeny usually dated by the Tournaisian age. Yet in the Tournaisian time, the general intensity of deformations, granitization, metamorphism, and island-arc volcanic activity sharply decreased everywhere except the West European regions, which experienced very active folding processes. In contrast, during the Famennian, folding influenced many mobil systems (that of the Urals, and the Innuitian, Appalachian, Peru-Bolivian, and others), being associated with higher levels of granitization and metamorphism.

The second period of generally activated tectonic processes lasted from the Visean until the end of the Permian. Like the previous one, it includes several epochs of gradually increasing tectonic activity, culminating in the Late Permian. The first Sudetic epoch of tectogenesis was in the Namurian. After the Early Carboniferous, intensive deformations were evolving in many structures of the Mediterranean, eastern Australia, and Cordillera, being accompanied by activated granitization. Island arc activity and ophiolite formation were regressing then, and this tendency lasted until the end of the Paleozoic. It reflected changes in the general style of tectonic processes, among which collision orogenesis became dominant, when continents were grouping into Pangea "II."

The Late Carboniferous Asturian orogeny was more energetic than the Sudetic phase. It was a period of main Variscan folding in the West European regions, as

well as in the Uralian and other systems. In addition, the extended processes of granitization of the terminal Carboniferous affected all mobil belts, being most intensive in the Mediterranean and Urals–Okhotsk systems.

The Saalic and Pfalzian epochs of tectogenesis, the final in the Paleozoic, evolved in the second half of the Permian period, with maxima in the Artinskian and Late Permian time. Before the Triassic, deformations spread over many mobil structures. Linear folding resulting from general compression was formed in the Equador-Colombian segment of Andes, and in the Cordillera, Ob'–Zaisan, and Gobi–Hinggan systems. In the Late Permian, granitization was at its maximum level detectable in the Paleozoic, and metamorphic processes were also very active. This suggests a predominance of compressive environments and collision-orogenic regimes that corresponds well to the inferred reconstruction of Pangea "II" before the end of the Paleozoic. The Late Permian is definitely a global epoch of tectogenesis, despite a sharp decrease of any volcanic activity.

After the Pfalzian orogeny, general tectonic activity was slowing down until the Late Triassic. The Hercynian era came to an end. Granitization and metamorphism were at a low activity level. Environments of extension and sea–floor spreading dominated in all mobil belts, as is suggested by the Middle Triassic peak of ophiolite formation. Simultaneously, this moment was the beginning of the new Kimmerian era. Its Early Kimmerian tectonic epoch is confined to the terminal Late Triassic interval and shows contrasting features. Riftogenesis, environments of extension, and regenerated mobil regimes were under progress in the western part of the Mediterranean belt. Yet, at the same time, intensive folding was evolving in its eastern regions. Folding and other deformations were active in the Caucasus–Kopet Dag, Lesser Caucasus–Iran, Pamirs, Hindukush, Himalayas, Kunlun Shan, Qinling, and Yunnan–Malayas systems, in addition to the northern regions of Cordillera system. These events are responsible for the large peak of tectonic activity on the histogram that precisely coincides in time with the distinct peak of island-arc volcanic activity and with the certain intensification of granitization and metamorphic events.

Afterward, there was the long Early-Middle Jurassic interval of low tectonic activity. We already pointed to the peculiar and enigmatic features of this interval, when magmatic, metamorphic, and tectonic activity was at a minimum level. Surprisingly, immense eruptions of continental traps occurred in southern Africa, Antarctica, Tasmania, and Patagonia during this interval (Fig. 3). The end of the Middle Jurassic marks a certain intensification of deformations in the Antarctic, the Andes, and in the Caucasus–Kopet Dag, Anyui–Chukchi, Amur–Okhotsk, and Verkhoyansk systems.

The third period of widespread tectonic activity begins with the Late Kimmerian epoch and culminates in the terminal Late Jurassic time. In this epoch, 16 systems of mobil belts experienced variable deformations of different intensities. Therefore, the Late Kimmerian events were definitely of a global rank. This epoch completed the Kimmerian tectonic era. Corresponding deformations were most effective in the structures framing the Pacific (the Verkhyansk–Kolyma regions, Japan, New Zealand, Cordillera, and the Andes). Again, the peak of these deformations correlates with peaks of such processes as island-arc volcanism, metamorphism, and granitization. Since the Late Jurassic there was the growing activity of ophiolite formation, but its maximum lay ahead in the Neocomian interval. The Late Kimmerian epoch of tectogenesis is the only one in the Mesozoic–Cenozoic era that was not preceded by a peak activity of sea–floor spreading processes. Still, global environments of extension existed even then, as is evident from the mentioned widespread eruptions of plateau basalts.

The spreading phase immediately followed the Late Jurassic stage in the Neocomian interval and lasted through the Hauterivian and Barremian. As was noted, the Neocomian was a period of highly intensive ophiolite formation. In addition, trap eruptions were active in the Parana region, southern Africa, and Pajmahal province. Simultaneously, processes of deformations, island-arc volcanism, and metamorphism were slowing down, which was followed by a decline of granitization in the Hauterivian–Barremian time. These events marked the beginning of the Alpine tectonic era; yet, it is enigmatic as to why they were associated with growing volcanic activity in marginal belts.

The Austrian epoch of tectogenesis of the Aptian–Albian terminated the Late Jurassic–Cretaceous period of high tectonic activity. The main deformations of this epoch were again concentrated in the marginal structures of the Pacific ocean (the South Anyui, Oloi, Sikhote Alin', Nippon, Taiwan–Philippines, Amur–Okhotsk, Kataysia–South China, Alaska, Cordillera, Antilles–Caribbean, and North Andean regions). They also affected the Mediterranean belt (the Apennines, Atlas, Carpatho-Balkan, Lesser Caucasus-Iranean, Yunnan–Malayas, and Indo-Burman systems). The culmination of thrust and fold deformations correlates with the peak activity of island-arc volcanism. Granitization, metamorphism, and volcanism in continental belts were also at their peak intensity during the Austrian epoch, and only ophiolite formation shows an opposite evolutionary trend. Undoubtedly, it was a global epoch of compression, followed by a decrease of tectonic activity

Another Laramide epoch of tectogenesis corresponds to the terminal Cretaceous–Paleocene interval, when tectonic events were controversial. The epoch is well depicted by deformations in the Carpatho-Balkan, Koryak–western Kamchatka, Olyutor–eastern Kamchatka, Skhalin–Hokkaido, Nippon, Taiwan–Philippines, and Alaskan systems. Yet, the widespread volcanic activity of island arcs showed a tendency of decreasing. Metamorphic processes were under rapid progress, but granitization was approaching a minimum. Ophiolite formation nearly ceased, which is typical of the compressive environments unfavorable for simultaneous basaltic eruptions on the Deccan Plateau.

In addition, the volcanic activity of continental marginal belts was also rapidly increasing. Nevertheless, the Laramide epoch of tectogenesis seems valid despite some difficulties in proving this, as we do not have necessary data on events within continental marginal zones that are now covered with shelf seas.

This also relates to the following Cenozoic epochs of tectogenesis. The Pyrenean (Inca) epoch, which originated during the terminal Eocene time, is the beginning of the *fourth period* of high tectonic activity. During this time, intensive thrust and fold deformations affected the Alpine, Dinaro-Anatolian, and Lesser Caucasus-Iranean systems. Later, the orogenic regime spread over these and nearby regions. The Inca folding is manifested all along the Andes and Cordillera. This peak of deformations coincides with a period of high volcanic activity of island arcs and continental marginal belts; in contrast, ophiolite formation was at a low level. Data on granitization and metamorphism is doubtful, as these events were poorly studied.

Environments of extension dominated during the Oligocene. Characteristic events of this stage include: back-arc spreading, ophiolite formation, basaltic eruptions in Patagonia, and limited occurrences of deformations and island-arc volcanism.

The Attic orogeny was initiated in the late Miocene and presumably lasted through the Pliocene and Pleistocene. It marks the last Cenozoic epoch of tectonic activity, which was more energetic than the preceding epoch. Its deformations mainly influenced the Mediterranean belt, from the Apennines on the west to the Indo-Burman and Sunda regions on the east. Also, Miocene–Pliocene folding was recognized in eastern Kamchatka, the Philippines, New Zealand, New Guinea, and the Andes. This epoch is characterized by extreme volcanic activity of island arcs and widespread regional metamorphism. As in the other epochs of high tectonic activity, ophiolite formation was insignificant. The Attic epoch is assumed to be continuing up to the present, i.e., the Earth's crust should now be under generally dominating stress conditions. This state of the Earth surface in recent time was earlier suggested by Kropotkin and other investigators.

It is remarkable that the Phanerozoic history of the Earth's tectonic evolution was periodic. Usually, epochs of tectogenesis were separated by intervals of 35 - 40 Ma, and this corresponds to cycles recognized in sedimentary sequences by many researchers. Their antiphases were commonly marked by periods of intensive ophiolite formation and trap eruptions characteristic of crustal extension.

In total, there were 16 Phanerozoic epochs of tectogenesis (Fig. 5). All of them except the Grampian (Finnmark, Sardic, or Delamerian) epoch of the Early Ordovician were different from the Middle Cambrian Salair epoch. Nearly all epochs were of global significance, as coeval events occurred in widely spaced regions. Still, none of the epochs simultaneously activated all mobil belts, or even all segments of a belt. To be correct, we must add that remote manifestations of particular phases were not, in a strict sense synchronous.

The periodicity of a lower rank is represented by the alternation of longer periods of intensive and quiet tectonic activity. Their history was not monotonous but disturbed by subordinate epochs of tectogenesis. The active periods of this rank correspond to the Caledonian (Late Ordovician–Devonian), Hercynian (Late Carboniferous–Permian), Kimmerian (Late Jurassic–Early Cretaceous), and Alpine (Eocene–Neogene) tectonic cycles. "Main" epochs of tectogenesis – the Ardennian of the Late Silurian, the Pfalzian of the Late Permian, the Late Kimmerian of the Late Jurassic, and the Attic of the late Miocene – were terminal in the corresponding cycles. These, i.e., Bertrand's cycles, lasted approximately 150 Ma each.

CONCLUSION

To correlate all secular variations of the geological processes under discussion, we used the synoptic scheme of global Phanerozoic events (Fig. 6). In addition to the considered histograms and variation curves it includes cumulative histograms of endogenic activity variations in mobil belts and platform regions and data on global transgression–regression periods and sea level fluctuation.

Our analysis shows that Phanerozoic processes of tectonic deformations, granitization, metamorphism, and volcanism were of variable intensity but continuously occurring in some region of the Earth. There was a general synchronization of the peak activity of deformations, regional metamorphism, island-arc and marginal belt volcanism, and, usually, granitization. It is conceivable that these processes are evidence of high endogenic activity and stress conditions, peaks of which were responsible for epochs of tectogenesis.

The cumulative histogram of endogenic activity variations in mobil belts shows the arithmetic mean values estimated by using the six other histograms to its left (see Fig. 6). There is no way to correctly evaluate the activity of the six geological processes under consideration, and we understand that this cumulative histogram is conventional. Still, it proves the validity of the global orogenic phases recognized by Stille, and of four tectonic cycles, three of which have been known since Bertrand's study. Hence, we believe that this histogram adequately depicts the endogenic "pulse of the Earth."

The above cycles are, in turn, parts of megacycles (Wilson's cycles), estimated to be approximately 600 Ma long. Two such megacycles have been distinguished in the Late Precambrian–Phanerozoic history of the Earth. The first of them lasted from the existence time of Pangea "I" until the appearance of Pangea "II." This megacycle, named after Wegener, includes the Baikalian, Caledonian, and Hercynian cycles. Its initial part corresponds to the period of the Pangea "I" breakup and its second part, to the period when its fragments, with newly formed crustal zones, were grouping into Pangea "II" by the end of the Paleozoic. The sec-

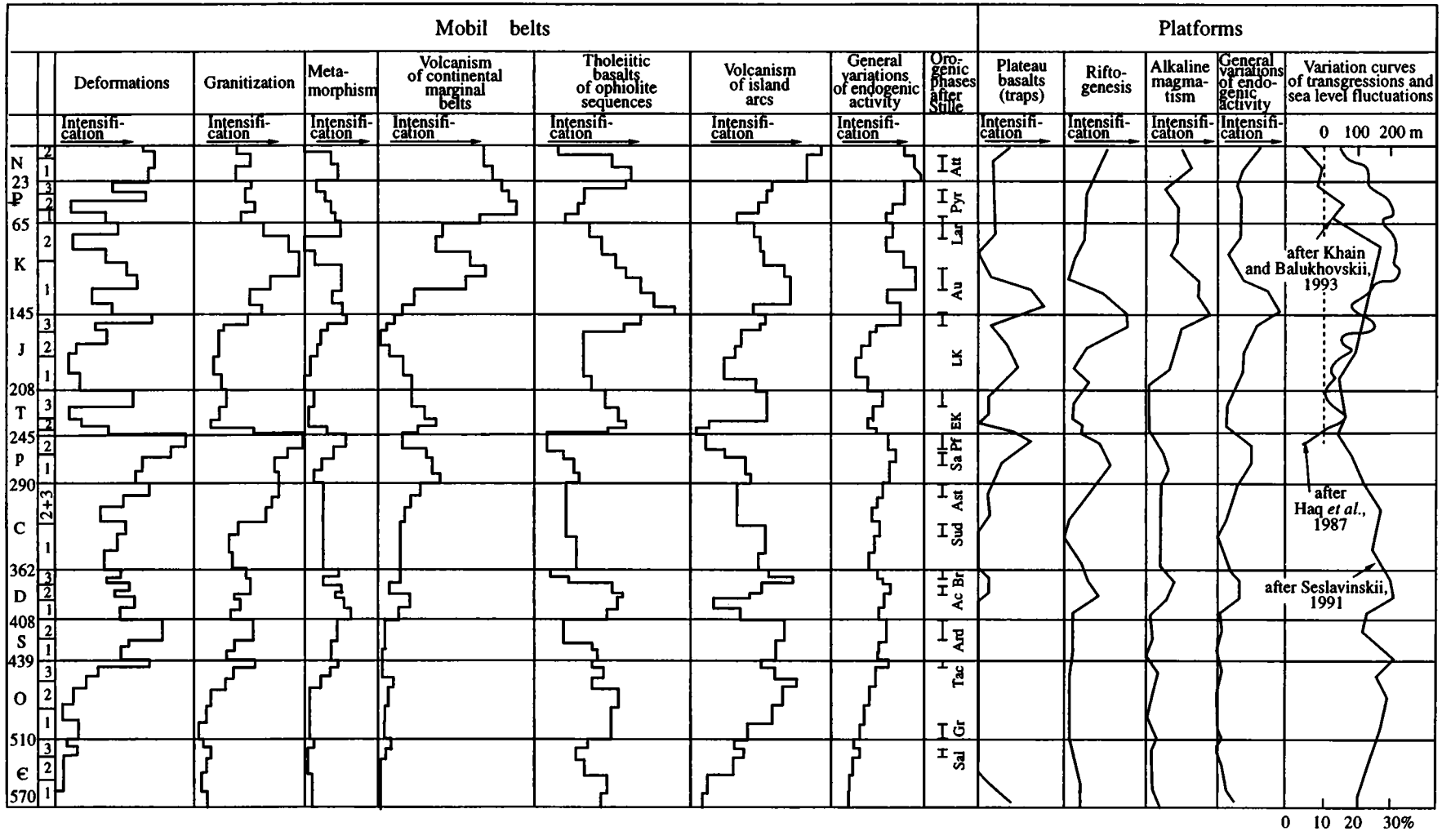

Fig. 6. Global correlation of tectonic, magmatic, and metamorphic events of the Phanerozoic in mobil belts and platform areas.

Orogenic phases after Stille: Sal – Salair; Gr – Grampian; Tac – Taconian; Ard – Ardennian; Ac – Acadian; Br– Bretonian; Sud– Sudetic; Ast – Asturian; Sa – Saalic; Pf – Pfalzian; EK – Early Kimmerian; LK – Late Kimmerian; Au – Austrian; Lar – Laramide; Pyr – Pyrenean; Att –Attic.

ond megacycle only went through its first Mesozoic–Cenozoic period of the Pangea "II" breakup. It includes the Kimmerian and Alpine cycles.

Because we are concerned in this work with a part of geological history limited by the Vendian–Cambrian boundary, the Baikalian (or Cadomian, Brazilian, Panafrican) cycle is not considered. Yet, it should be noted that the Baikalian, Caledonian, and Hercynian cycles, as elements of the Late Precambrian–Paleozoic megacycle, were of different rank and significance. The Baikalian cycle was preliminary but nevertheless resulted in the consolidation of Gondwana as a whole, with the formation of a vast epi-Baikalian platform (partly destructed later) in the western, northern, and northeastern marginal zones. In addition, the similar South Barents–Pechora platform was formed far on the north.

The Caledonian cycle is central in the megacycle. A significant destruction of continental crust (abundant ophiolites as indicators) was in action during its initial stage, while the mountain building, granitization, and first Phanerozoic generation of volcano-plutonic marginal belts were characteristic of the terminal period of the cycle. The peak of island-arc volcanic activity is confined to the Caledonian cycle, with a short prolongation into the very beginning of the Hercynian cycle. From the end of the Caledonian cycle (the Devonian time), widespread mountain building processes started to affect the older platform areas adjacent to orogens.

Respectively, the Hercynian cycle is terminal in the megacycle (according to Shatskii, it is complementary to the Caledonian cycle). The oceanic crust formation (in particular structures only) of its initial stage is a continuation of corresponding processes started in the Ordovician, Silurian, or Early Devonian, i.e., in the Caledonian cycle. Deformations, regional metamorphism, granite formation, and orogenesis (molasses as indicators) were at their maxima in the Permian. In the second half of the Hercynian cycle, there was initiated another stage of oceanic crust formation, which terminated with the Mesozoic folding and orogenesis. Respectively, we are dealing again with the time overlapping of initial and terminal stages of adjacent cycles.

The Kimmerian cycle was opening the new megacycle of the Earth's evolution. Its beginning in the middle Triassic was signified by widespread processes of destruction of the continental crust, namely by riftogenesis and ophiolite formation. In some regions, these processes had already been activated in the terminal Paleozoic time. The maximum of tectonic activity was confined to the end of the Late Jurassic, i.e., to the Late Kimmerian tectonic epoch. This was a period of widespread deformations, island-arc volcanism, metamorphism, orogenic events, and granitization, which were terminating the Kimmerian epoch.

The Alpine cycle also begins with the intensification of sea–floor spreading in the Late Jurassic, i.e., again in the previous cycle. The peak of spreading activity was in the Neocomian, when other tectonic processes were still at a low level. Then there were several tectonic epochs of gradually growing general activity, which finally reached a maximum in the last Attic epoch of the late Miocene–Pliocene time. The formation of modern rugged topography, extremely active island-arc volcanism, widespread metamorphism, but low ophiolite formation rate, i.e., all the features of an epoch terminating the tectonic era, were characteristic of this time.

The cumulative variation curve for the tectono-magmatic events in platform areas provides a summary of data of the three left variation curves (see Fig. 6) and distinctly reveals the pulsation of endogenic activity. Especially remarkable is the time coincidence of two large active periods, the Permian and Late Jurassic–Neocomian, with the two most prominent tectonic epochs in mobil belts, the Pfalzian and Late Kimmerian, respectively.

Figure 6 also shows the secular variations of sea areas periodically spreading over continents. The paleozoic part of the curve was published by Seslavinskii (1991), who described the evaluation technique. The Mesozoic–Cenozoic part was obtained in the same manner (Khain and Balukhovskii, 1993). Main events of sea transgression and regression, distinguishable on the Mesozoic–Cenozoic curve, are well correlated with the corresponding global fluctuations of sea level studied by Haq *et al.*, (1987). Similar results obtained by different methods suggest that the Paleozoic part of our curve is also reliable, and thus both curves can be used to analyze the paleogeography of the Earth. The worldwide regressions of the Paleozoic were distinctly confined to the terminal periods of the Caledonian and Hercynian cycles. Unfortunately, the less detailed data on the Cretaceous period does not allow us to analyze the paleogeography of the boundary period between the Kimmerian and Alpine cycles.

Thus, the four recognized Phanerozoic periods of particularly high endogenic activity of the Earth – the Late Silurian, Late Permian, Late Jurassic, and Late Miocene–Pliocene – were terminal in the Caledonian, Hercynian, Kimmerian, and Alpine cycles, respectively. The cycles do not reveal an absolute periodicity, and lasted from 120 to 150 Ma. The cycle of the Alpine era has already lasting for approximately 140 Ma. It is now near its culmination, which seems to be 5 - 10 Ma ahead if the inferred periodicity of tectonic processes is valid.

The megacycles of the Earth's history, or rather the large periods of the formation and breakup of supercontinents (Pangea "I" and Pangea "II") bearing the entire mass of continental crust created at those times were presumably caused by an essential rearrangement of the mantle convection system, namely by the transition from the two-cell to one-cell convection that would explain the megacycles. As for Bertrand's cycles, they probably were related to changes in convection patterns only in the upper mantle. Reorganizations of a lesser rank in plate kinematics were probably responsible for the periodic activity of tectonic deformations, granite magmatism, and regional metamorphism. It is clear

enough that these three-rank cycles should also be related to variations in the intensity of heat–flow from the Earth's interior. Longer periods of heat accumulation should be alternating with shorter phases of heat release at the rearrangement stages of mantle convection. For example, heat accumulation under supercontinents with thick crustal and lithospheric layers of low conductivity could result in their breakup. Such a mechanism has already been discussed in published works.

Reviewer M.A. Semikhatov

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Sea Transgressions and Regressions on Continents during the Phanerozoic (a Quantitative Analysis)[1]

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Received December 22, 1992

Abstract – The focus of this study was to measure sea areas, which were covering continents, and to compare amplitudes of transgressions and regressions within platform, geosynclinal, and orogenic belt zones of all continents (except Antarctica) in the historical succession of the Phanerozoic epochs since the Early Cambrian to the Pliocene. I also estimated the numerical ratios between areas of marine and continental sedimentation at different evolutionary stages of these global structures. Finally, I sought to propose reasons for sea ingression upon continents and to understand how the transgressions and regressions are related to vertical movements of the earth crust, sea level fluctuations, and glaciations. Similar problems were recently considered by Seslavinskii (1991), who made an attempt to evaluate the development of global transgressions and regressions for the Paleozoic only.

To solve the problems under study, sea and land areas were measured, and volumes of most significant sedimentary formations were calculated using a series of litho-paleogeographic maps of the world. Ronov, Khain, Seslavinskii, and Balukhovskii (Ronov *et al.*, 1984; 1989) compiled these maps on the basis of available data on geology, lithologic composition, and paleogeographic environments of sediment formation in main structural zones (platforms, geosynclines, orogenic belts) of all continents. The maps correspond to 28 stratigraphic intervals of the Phanerozoic epochs, characterized by data on many thousands of natural and drilled sections in various regions of the Earth.

Measurement results for land, sea, and sedimentation areas in each structural zone of the continents are represented in Table 1. In historical succession, Table 1 also shows percentages of areas covered by sea relative to the bulk areas of corresponding tectonic zones.

For the maps used, an error of sea and land area measurements is ±10% on average. However, the estimated values become less confident from the top down to lower units of the stratigraphic scale, as well as when passing on a separate map from platforms to geosynclines or from the northern to the southern hemisphere. Presented diagrams are plotted using the latest version of the geological time scale (Harland *et al.*, 1990).

Analysis of the data obtained shows, first of all, that areas of sea ingression upon continents were regularly varying with time, once increasing, and then sharply decreasing during the tectonic cycles (Fig. 1). Respectively, there were "geocratic" (epeirocratic) and thalassocratic epochs in the Earth's history.

During the Quaternary and at present, the global ocean with inner and marginal seas covers 361 million km^2, i.e., 71% of the Earth's surface, which is the minimum area. Our epoch is geocratic, characterized by sea regression from elevated continents. Such epochs also existed in the past. They coincided with the final stages of tectonic cycles and are distinguished as been represented by: the Early Cambrian epoch at the boundary between the Baikalian and Caledonian cycles, the Late Silurian–Early Devonian period at the boundary between the Caledonian and Hercynian cycles, the Late Permian–Middle Triassic interval at the boundary between the Hercynian and Alpine cycles, and the Pliocene–Quaternary interval marking the boundary between the Alpine and a forthcoming cycle. The curve in Fig. 1 clearly shows that regression amplitudes were growing from cycle to cycle, attaining the current maximum.

During the thalassocratic epochs related to the middle stages of the cycles, areas of inland seas were growing significantly because of the continental subsidence and large transgressions. The Middle Ordovician, Middle Devonian, and Late Cretaceous intervals correspond to such thalassocratic epochs of the Phanerozoic. As is evident from Fig. 1, the extent of sea transgressions on continents was gradually decreasing from the Caledonian to Alpine cycle.

Thus, the results of area measurements suggest the first general regularity: sea areas on continents progressively decreased from the Early Paleozoic to the present epoch. For this period, the total contraction of sea areas attained 40 million km^2.

Comparing secular variations in sea area distribution, one may note another important and regular tendency: the largest sea transgressions and regressions were synchronous in both the platforms and geosyn-

[1] The lecture delivered at several universities in the USA (Cornell, Syracuse, Pennsylvania, Michigan, Yale, and Harvard) during the fall of 1991.

Table 1. Data on sea transgressions and regressions during the Phanerozoic for continents in bulk and for their main tectonic zones

Stratigraphic interval	Time span of the interval, 10^6 years (after Harland *et al.*, 1989; Odin, 1982; Palmer, 1983)	Main tectonic zones of continents	Areas, 10^6 km^2			Percentage of sea covered areas relative to the total area of a tectonic zone
			Erosion areas	Accumulation areas	Sea covered areas	
Pliocene (N_2)	5.2 – 1.6 = 3.6	Platforms	73.40	21.44	4.45	5
		Geosynclines	1.32	1.58	0.99	34
		Orogens	21.23	14.23	1.91	5
		Continents in bulk	95.95	37.25	7.35	5
Miocene (N_1)	35.4 – 23.5 = 11.9	Platforms	74.55	20.14	6.53	7
		Geosynclines	0.74	2.13	2.10	73
		Orogens	20.47	14.64	4.30	12
		Continents in bulk	95.76	36.91	12.93	10
Oligocene (P_3)	34.5 – 23.5 = 11.9	Platforms	77.80	17.71	5.85	6
		Geosynclines	1.76	3.91	3.58	63
		Orogens	21.26	9.23	2.76	9
		Continents in bulk	100.82	30.85	12.19	9
Eocene (P_2)	56.5 – 35.4 = 21.1	Platforms	75.35	24.31	14.67	15
		Geosynclines	2.27	7.95	7.18	70
		Orogens	14.73	8.00	2.19	10
		Continents in bulk	92.35	40.6	24.04	18
Paleocene (P_1)	65.0 – 56.5 = 8.5	Platforms	77.37	20.32	12.35	13
		Geosynclines	3.82	6.13	6.03	61
		Orogens	19.06	5.86	1.87	7
		Continents in bulk	100.20	32.31	20.25	15
Late Cretaceous (K_2)	97.0 – 65.0 = 32.0	Platforms	56.90	41.76	22.41	23
		Geosynclines	1.37	16.82	14.53	80
		Orogens	9.28	7.45	1.81	11
		Continents in bulk	67.55	66.03	38.72	29
Early Cretaceous (K_1)	145.6 – 97.0 = 48.6	Platforms	59.77	40.06	17.04	17
		Geosynclines	3.55	18.76	15.12	70
		Orogens	9.59	7.06	0.78	5
		Continents in bulk	72.91	65.88	33.54	24
Late Jurassic (J_3)	157.1 – 145.6 = 11.5	Platforms	74.30	25.15	16.82	17
		Geosynclines	4.91	16.05	14.10	67
		Orogens	8.13	4.12	1.44	12
		Continents in bulk	87.34	45.32	32.36	24
Middle Jurassic (J_2)	178 – 157 = 21	Platforms	75.08	25.47	14.83	15
		Geosynclines	2.29	12.76	12.49	83
		Orogens	11.54	6.04	1.40	8
		Continents in bulk	88.91	44.27	28.79	22
Early Jurassic (J_1)	208 – 178 = 30	Platforms	79.85	22.97	11.40	11
		Geosynclines	1.54	10.93	10.86	87
		Orogens	10.85	5.73	1.35	8
		Continents in bulk	92.24	39.63	23.61	18
Late Triassic (T_3)	235 – 208 = 27	Platforms	78.75	23.53	10.44	9
		Geosynclines	1.56	13.63	13.27	87
		Orogens	9.81	3.63	0.50	4
		Continents in bulk	90.12	40.79	24.21	18

Table 1. (Contd.)

Stratigraphic interval	Time span of the interval, 10^6 years (after Harland *et al.*, 1989; Odin, 1982; Palmer, 1983)	Main tectonic zones of continents	Areas, 10^6 km^2			Percentage of sea covered areas relative to the total area of a tectonic zone
			Erosion areas	Accumulation areas	Sea covered areas	
Middle Triassic (T_2)	241 – 235 = 6	Platforms	82.00	17.95	10.65	11
		Geosynclines	1.69	11.57	11.36	86
		Orogens	23.19	1.77	0.69	5
		Continents in bulk	96.89	31.29	22.70	18
Early Triassic (T_1)	245 – 241 = 4	Platforms	79.57	24.73	11.49	11
		Geosynclines	1.94	10.22	9.96	82
		Orogens	10.52	1.65	0.07	1
		Continents in bulk	92.03	36.60	21.52	17
Late Permian (P_2)	256 – 245 = 11	Platforms	75.61	19.34	11.55	12
		Geosynclines	1.07	9.23	8.94	86
		Orogens	14.58	4.88	1.81	9
		Continents in bulk	91.26	33.45	22.30	18
Early Permian (P_1)	290 – 256 = 34	Platforms	68.88	24.70	13.96	15
		Geosynclines	0.98	9.36	9.18	89
		Orogens	13.07	7.73	4.02	19
		Continents in bulk	82.93	41.79	27.16	22
Late and Middle Carboniferous (C_{2+3})	323 – 290 = 33	Platforms	68.26	28.04	16.69	17
		Geosynclines	1.02	9.66	9.46	89
		Orogens	11.53	8.81	5.36	26
		Continents in bulk	80.81	46.51	31.51	25
Early Carboniferous (C_1)	363 – 323 = 40	Platforms	71.91	20.99	15.92	17
		Geosynclines	2.38	17.63	17.18	86
		Orogens	8.67	4.03	2.36	19
		Continents in bulk	82.96	42.65	35.46	28
Late Devonian (D_3)	377 – 363 = 14	Platforms	67.65	25.39	21.80	23
		Geosynclines	2.50	18.55	17.99	85
		Orogens	7.32	4.10	2.23	19
		Continents in bulk	77.47	48.04	42.02	33
Middle Devonian (D_2)	386 – 377 = 9	Platforms	68.75	26.17	24.23	25
		Geosynclines	1.83	18.82	18.43	89
		Orogens	7.49	3.57	2.04	18
		Continents in bulk	78.17	48.56	44.70	35
Early Devonian (D_1)	408 – 386 = 22	Platforms	74.53	16.90	14.51	16
		Geosynclines	2.39	17.48	17.29	87
		Orogens	6.94	4.18	2.45	22
		Continents in bulk	83.86	38.56	34.25	28
Late Silurian (S_2)	424 – 409 = 15	Platforms	72.74	22.29	17.65	19
		Geosynclines	2.89	15.24	14.93	82
		Orogens	8.16	2.12	1.65	16
		Continents in bulk	83.79	39.65	34.23	28
Early Silurian (S_1)	439 – 424 = 15	Platforms	66.75	28.38	27.56	29
		Geosynclines	2.92	17.08	16.82	84
		Orogens	4.70	2.13	1.71	25
		Continents in bulk	74.37	47.59	46.09	38

Table 1. (Contd.)

Stratigraphic interval	Time span of the interval, 10^6 years (after Harland *et al.*, 1989; Odin, 1982; Palmer, 1983)	Main tectonic zones of continents	Areas, 10^6 km^2			Percentage of sea covered areas relative to the total area of a tectonic zone
			Erosion areas	Accumulation areas	Sea covered areas	
Early Cambrian (O_3)	464 – 439 = 25	Platforms	69.49	26.85	22.50	23
		Geosynclines	3.84	18.52	18.24	82
		Orogens	2.14	0.33	0.27	11
		Continents in bulk	75.47	45.70	41.01	34
Middle Ordovician (O_2)	476 – 464 = 12	Platforms	69.28	27.90	27.44	28
		Geosynclines	2.46	19.78	19.51	88
		Orogens	1.40	0.41	0.26	14
		Continents in bulk	73.14	48.09	47.21	39
Early Ordovician (O_1)	510 – 476 = 34	Platforms	70.26	25.40	24.65	26
		Geosynclines	2.38	18.32	18.32	88
		Orogens	1.21	0.46	0.38	23
		Continents in bulk	73.85	44.18	43.35	37
Late Cambrian ($\mathcal{E}_3$)	517 – 510 = 7	Platforms	66.54	24.43	22.93	25
		Geosynclines	2.67	15.61	15.46	84
		Orogens	6.02	0.92	0.46	7
		Continents in bulk	75.23	40.96	38.85	33
Middle Cambrian ($\mathcal{E}_2$)	536 – 517 = 19	Platforms	66.77	25.65	22.17	24
		Geosynclines	2.01	14.47	14.41	87
		Orogens	6.76	1.03	0.45	6
		Continents in bulk	75.54	41.15	37.03	32
Early Cambrian ($\mathcal{E}_1$)	570 – 536 = 34	Platforms	55.78	22.94	18.23	23
		Geosynclines	1.78	13.96	13.86	88
		Orogens	15.87	1.76	1.28	7
		Continents in bulk	73.43	38.66	33.37	30

Except Antarctica and areas with no data available.

clines, having comparable periods. This trend is clearly seen from the parallel fluctuations of corresponding curves in Fig. 2.[2] Seas of orogenic belts deviated from this regularity, and the transgressions had a different periodicity there, as is evident from the diagram.

Data in Table 2 shows that during the Phanerozoic, the Laurasian platforms (Russian, Siberian, Chinese, and North American) were generally different from the Gondwanan platforms (African, South American, Australian, and Indian) in having larger sea areas and a higher extent of transgressions combined with lesser amplitudes of regressions (Fig. 3). Omitting minor details, one can see that in both groups of continents, these events were almost synchronous during the Alpine and Hercynian cycles and show different trends only within the Caledonian cycle.

Comparing secular variations in the sea area distribution for the largest individual ancient platforms (Fig. 4), we can recognize the time diversity in development of transgressions and regressions, which was first noted by Yanshin (1973). However, the above data shows that for the platforms and geosynclines in total, these events had a universal character (see Fig. 2). This suggests that they are dependent on global processes, periodically initiating or reducing the tectonic activity that is indirectly manifested in transgressions and regressions of the global ocean. This global rhythm is evident despite the variable movements of separate large blocks of the lithosphere, noted by Yanshin. This shows the global tendency to be dominant over regional fluctuations (Ronov *et al.*, 1976).

To continue our analysis, let's consider the time distribution of total continental area between inland seas and land provinces taking for them the provinces of continental sedimentation (Fig. 5). This diagram shows that during the Phanerozoic, the amount of continental surface covered with seas ranged between 5 and 40% of the total area.

Therefore, we come to the next general conclusion: once expanding and then contracting, land environments were dominant on continents during all epochs

[2] For stratigraphic intervals shorter than an epoch, there was sometimes a minor delay in initiation of transgressions on platforms relative to adjacent geosynclines, and vice versa, the earlier beginning of regressions in the latter zones.

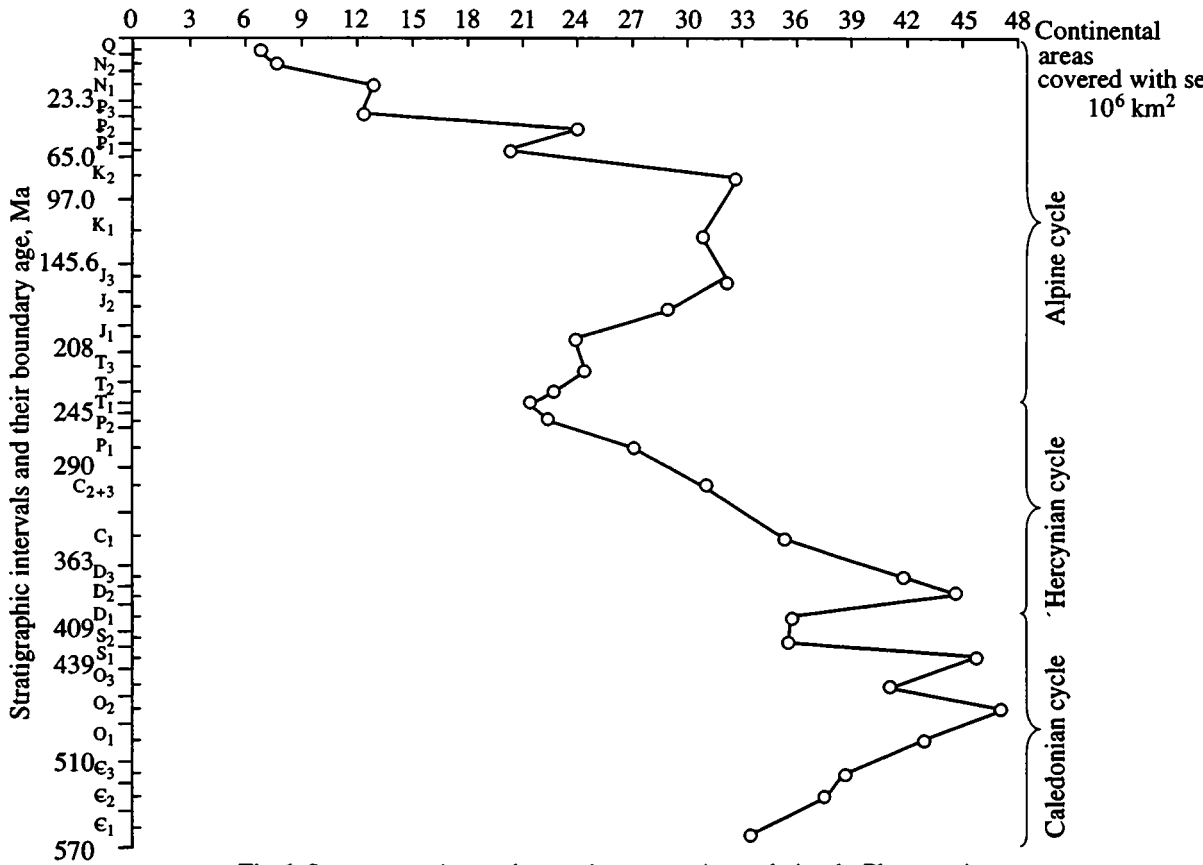

Fig. 1. Sea transgressions and regressions on continents during the Phanerozoic.

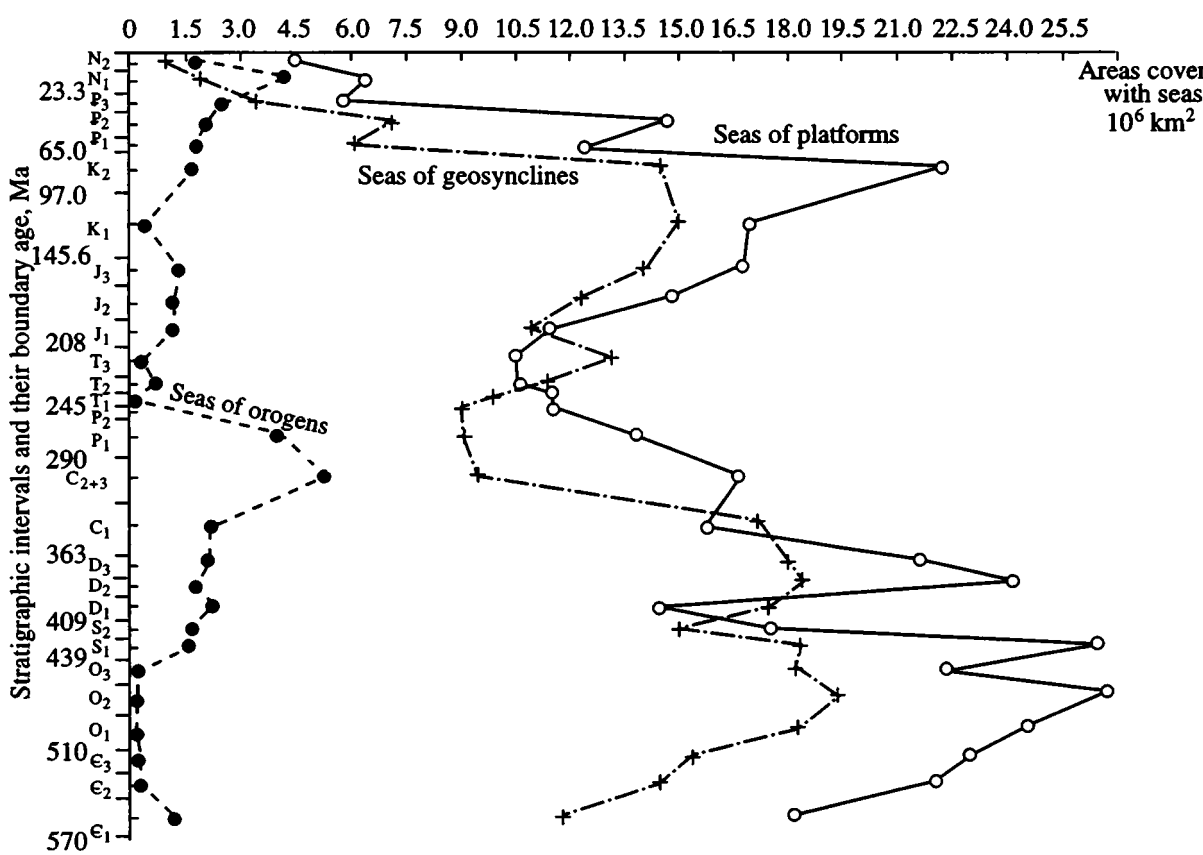

Fig. 2. Sea transgressions and regressions within continental platforms, geosynclines, and orogens during the Phanerozoic.

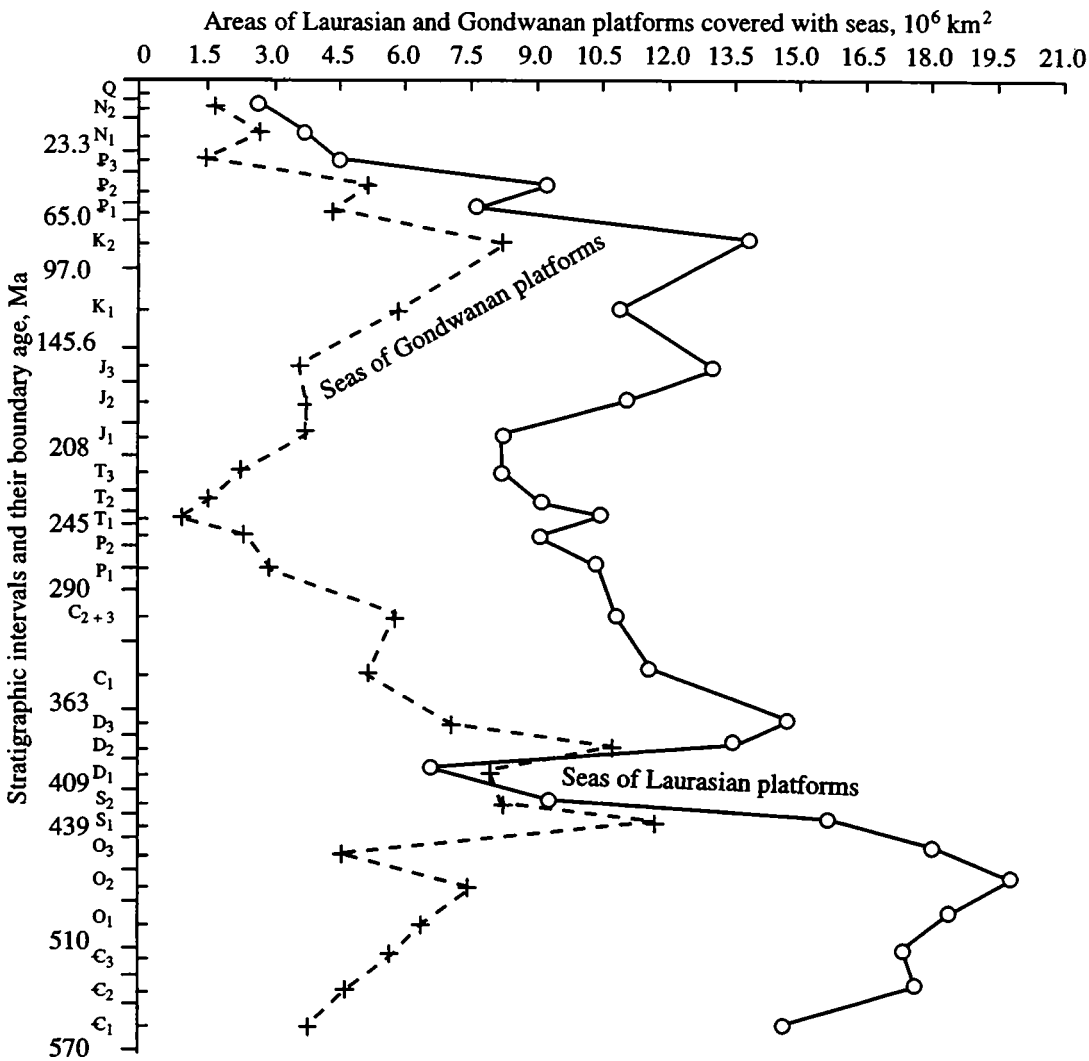

Fig. 3. Sea transgressions and regressions on the Laurasian and Gondwanan platform groups during the Phanerozoic.

of the Phanerozoic. Against this background, there is a clear tendency of land growth since the Cambrian until the present. Stable and mobil continental zones differed sharply in their proportions of sea and land areas. The platforms (Fig. 6) always zones of widespread dry lands, and seas covered only 5 to 30% of their total area. On the contrary, seas dominated the geosynclines, covering up to 90% of their entire area during the Paleozoic (Fig. 7). Only in the end of the Mesozoic and during the Cenozoic (especially in the Pliocene) were the sea areas cardinally reduced here. Seas were very insignificant in orogenic belts, and had a somewhat larger total area during the Paleozoic as compared to the Mesozoic–Cenozoic period (Fig. 8). Thus, during all stages of the Phanerozoic history, the geosynclines, which represent zones of maximum amplitude and the rate of the Earth crust subsidence (Ronov, 1980, 1993; Ronov *et al.*, 1990), were not only areas of maximum sea expansions, but also the natural ways for sea water migration onto continents.

According to our data on secular variations in relative proportions of the marine and terrestrial sedimentation areas on continents and within their main structural zones (Table 1), there is another empiric regularity. Despite the periodic changes in sedimentation environments and relative volumes of marine and continental sediments, there were evidently irreversible and geochronologically progressive changes in the paleogeographic settings of sediment accumulation (Ronov, 1993).

For continents in bulk, they are manifested in a progressive areal expansion of sediments accumulated under terrestrial conditions since the Cambrian until recent time, as well as the corresponding reduction of marine sedimentation areas (Fig. 9). This trend is most obvious for the platforms (Fig. 10). For orogenic belts it is less evident (Fig. 11), being hardly visible in the geosyncline zones of the Alpine cycle (Fig. 12).

The considered numerical evaluations at the global rank show that planetary transgressions and regressions

Table 2. Data on sea transgressions and regressions during the Phanerozoic for platforms of the Laurasian and Gondwanan groups

Stratigraphic interval	Time span of the interval, 10^6 years (after Harland *et al.*, 1989)	Group of platforms	Areas, 10^6 km^2			Percentage of sea covered areas relative to the total area of the Laurasian or Gondwanan group of platforms
			Erosion areas	Accumulation areas	Sea covered areas	
Pliocene (N_2)	5.2 – 1.6 = 3.6	Laurasian	31.80	8.63	2.69	7
		Gondwanan	41.60	12.81	1.76	3
Miocene (N_1)	23.5 – 5.2 = 18.3	Laurasian	32.14	7.93	3.79	9
		Gondwanan	42.41	12.21	2.74	5
Oligocene (P_3)	35.4 – 23.5 = 11.9	Laurasian	30.92	9.20	4.42	11
		Gondwanan	46.88	8.51	1.43	3
Eocene (P_2)	56.5 – 35.4 = 21.1	Laurasian	32.51	11.93	9.42	21
		Gondwanan	42.84	12.38	5.25	9
Paleocene (P_1)	65.0 – 56.5 = 8.5	Laurasian	32.22	9.89	7.81	18
		Gondwanan	45.10	10.43	4.54	8
Late Cretaceous (K_2)	97.0 – 65.0 = 32.0	Laurasian	23.16	31.64	14.04	26
		Gondwanan	33.74	10.12	8.37	19
Early Cretaceous (K_1)	145.6 – 97.0 = 48.6	Laurasian	28.65	19.91	10.91	23
		Gondwanan	31.13	20.54	6.13	12
Late Jurassic (J_3)	157.1 – 145.6 = 11.5	Laurasian	27.35	16.5	13.17	30
		Gondwanan	46.95	8.90	3.65	6
Middle Jurassic (J_2)	178 – 157 = 21	Laurasian	27.17	17.48	11.11	25
		Gondwanan	47.91	7.99	3.72	7
Early Jurassic (J_1)	208 – 178 = 30	Laurasian	31.48	15.58	8.29	18
		Gondwanan	48.37	7.38	3.11	6
Late Triassic (T_3)	235 – 208 = 27	Laurasian	30.71	16.06	8.24	18
		Gondwanan	48.04	7.47	2.20	4
Middle Triassic (T_2)	241 – 235 = 6	Laurasian	33.87	12.50	9.17	20
		Gondwanan	48.14	5.45	1.48	3
Early Triassic (T_1)	245 – 241 = 4	Laurasian	31.43	18.30	10.60	21
		Gondwanan	48.14	6.43	0.89	2
Late Permian (P_2)	256 – 245 = 11	Laurasian	30.21	13.12	9.03	21
		Gondwanan	45.40	6.22	2.52	5
Early Permian (P_1)	290 – 256 = 34	Laurasian	18.37	13.73	10.47	33
		Gondwanan	40.51	10.97	3.49	7
Late and Middle Carboniferous (C_{2+3})	323 – 290 = 33	Laurasian	27.03	14.42	10.86	26
		Gondwanan	41.23	13.58	5.83	11
Early Carboniferous (C_1)	363 – 323 = 40	Laurasian	26.05	13.56	11.53	29
		Gondwanan	45.86	7.43	4.39	8
Late Devonian (D_3)	377 – 363 = 14	Laurasian	24.96	15.85	14.76	36
		Gondwanan	42.69	9.54	7.04	13
Middle Devonian (D_2)	386 – 377 = 9	Laurasian	26.96	14.66	13.48	32
		Gondwanan	41.79	11.51	10.75	20
Early Devonian (D_1)	409 – 386 = 23	Laurasian	31.99	7.73	6.65	17
		Gondwanan	42.54	9.17	7.86	15
Late Silurian (S_2)	424 – 409 = 15	Laurasian	30.85	10.71	9.35	22
		Gondwanan	41.89	11.58	8.30	15
Early Silurian (S_1)	439 – 424 = 15	Laurasian	26.04	16.26	15.81	37
		Gondwanan	40.71	12.12	11.75	22
Late Ordovician (O_3)	464 – 439 = 25	Laurasian	25.09	18.35	17.90	41
		Gondwanan	44.40	8.50	4.60	9
Middle Ordovician (O_2)	476 – 464 = 12	Laurasian	24.89	20.06	19.90	44
		Gondwanan	44.39	7.84	7.54	14
Early Ordovician (O_1)	510 – 476 = 34	Laurasian	24.74	18.41	18.41	43
		Gondwanan	45.52	6.99	6.24	12
Late Cambrian ($\mathcal{E}_3$)	517 – 510 = 7	Laurasian	24.53	18.27	17.36	41
		Gondwanan	42.0	6.16	5.57	12
Middle Cambrian ($\mathcal{E}_2$)	536 – 517 = 19	Laurasian	22.73	19.90	17.54	41
		Gondwanan	44.04	5.75	4.63	6
Early Cambrian ($\mathcal{E}_1$)	570 – 536 = 34	Laurasian	22.00	17.37	14.56	37
		Gondwanan	33.29	5.57	3.67	9

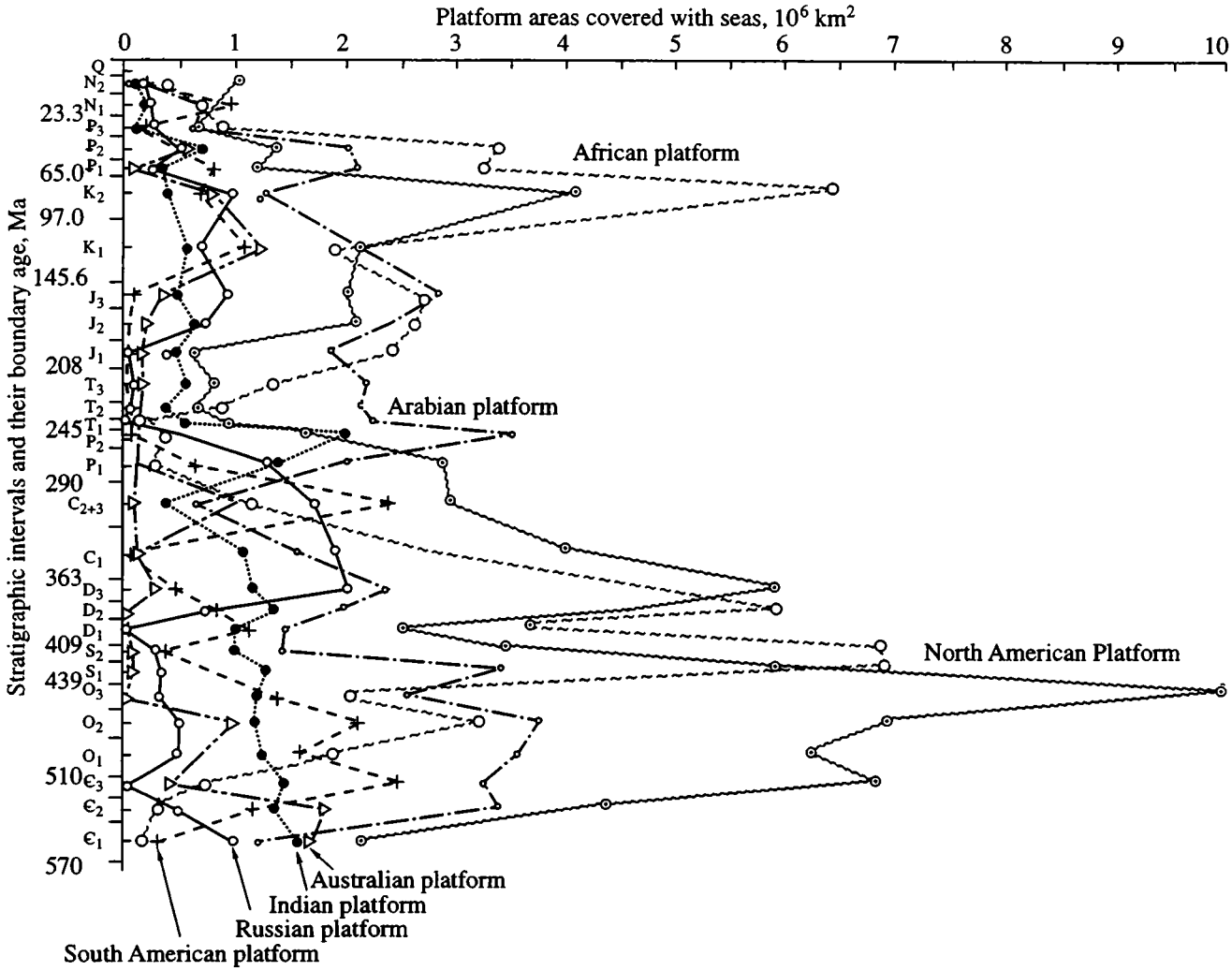

Fig. 4. Sea transgressions and regressions on selected ancient platforms of continents during the Phanerozoic.

were real phenomena and were confined to certain stages of the Caledonian, Hercynian, and Alpine tectonic cycles of the Earth's history.

It is remarkable that our curve for the inland sea area variations (Fig. 13) is similar to the curve after Vail *et al.* (1977), depicting the eustatic sea level fluctuations during the Phanerozoic. The correlation of both curves shows that an expansion of sea areas corresponds to sea level elevation above the modern position, while their reduction coincides with the sea level lowering. This coincidence of two curves plotted by different independent methods suggests that sea level fluctuations and marine transgressions and regressions on continents are of common origin. There is no doubt that being coeval, these processes are related to the Earth crust subsidence and uprise in the continental and oceanic segments.

In Fig. 14, the transgression–regression curve is correlated with the curve integrating average values of the continental subsidence or uprise at the various Phanerozoic epochs (Ronov *et al.*, 1976). Naturally, one should expect an increase of sea areas when continental subsidence becomes dominant over their uprise, and vice versa, their decrease when the subsidence has a lesser significance than uplifts. Yet, such correlations were not always observable, as is evident from the diagram. This suggests that there were other factors in action. Undoubtedly, the global glaciations operated as one of them, consolidating huge water masses inside the polar provinces. In fact, the arrows in Fig. 14, which mark global glaciations of the Pliocene, Late Permian, and Late Silurian time, do coincide with the maximum sea regressions.

Donovan and Jones (1979) believe that sea level fluctuations are related to changes in the ocean floor morphology, i.e., to the capacity of oceanic basins affected by volumetric changes of mid-ocean ridges.

From this viewpoint, oceanic history is traceable only since the Late Jurassic. But so far, there is no published data that prove this hypothesis as applicable for the Paleozoic and Early Mesozoic epochs.

Thus, the real causes of sea transgressions and regressions on continents are still not well understood.

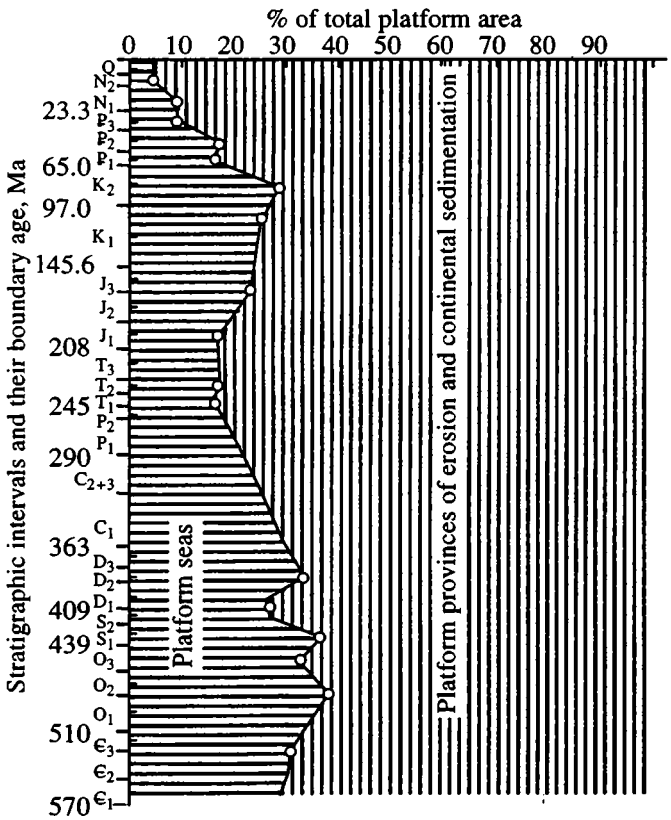

Fig. 5. Secular variations in areal proportions of inland seas relative to continental land with provinces of continental sedimentation.

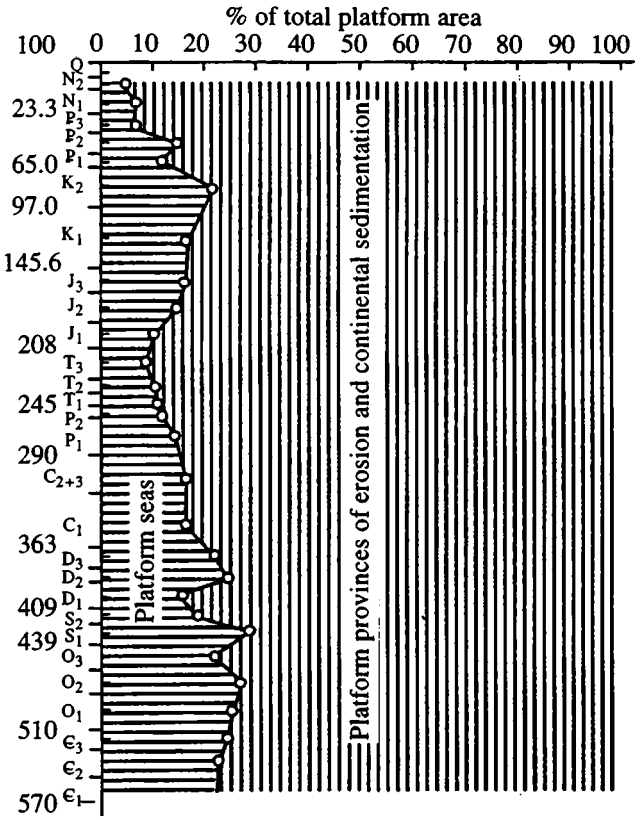

Fig. 6. Secular variations in areal proportions between seas and land with provinces of continental sedimentation for continental platforms.

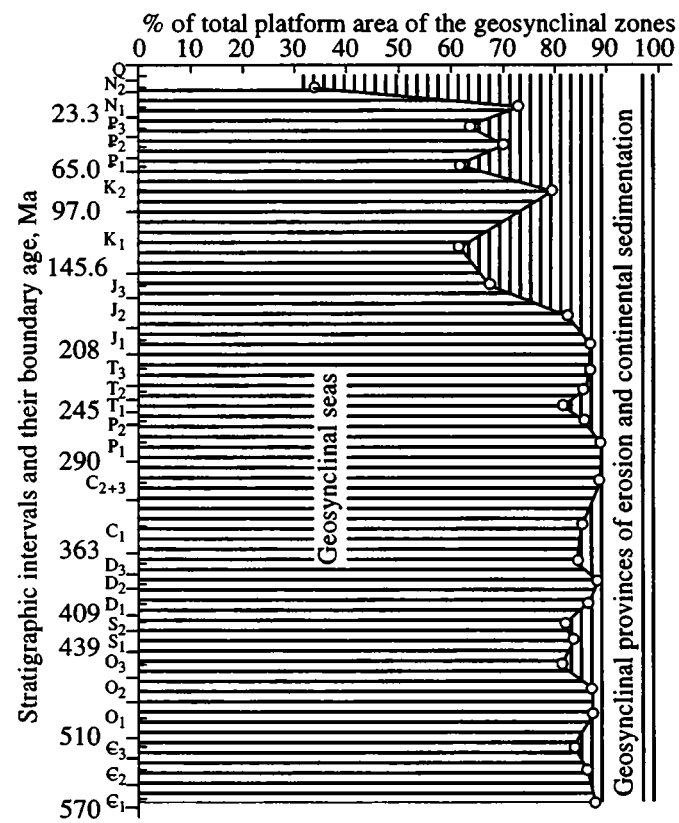

Fig. 7. Secular variations in areal proportions between seas and land with provinces of continental sedimentation for geosynclinal zones.

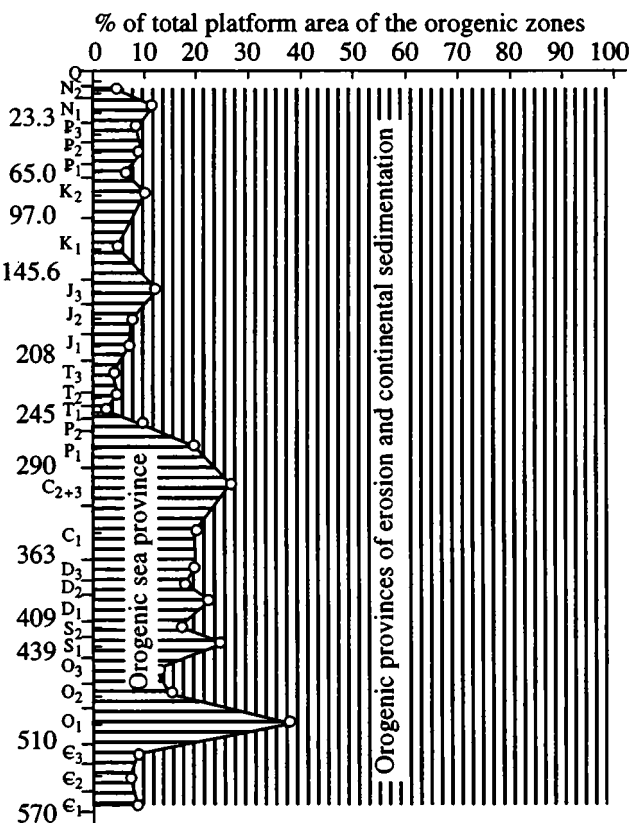

Fig. 8. Secular variations in areal proportions between seas and land with provinces of continental sedimentation for continental orogenic zones.

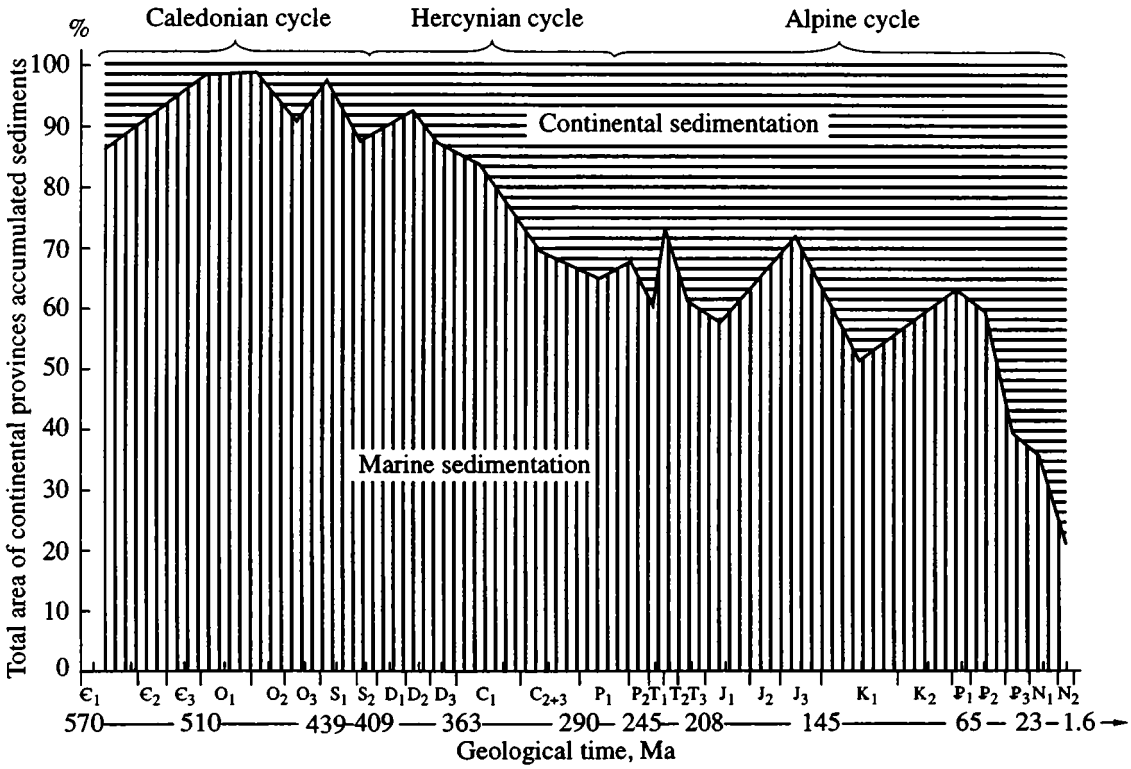

Fig. 9. Areal variations of marine and continental sedimentary provinces of continents during the Phanerozoic.

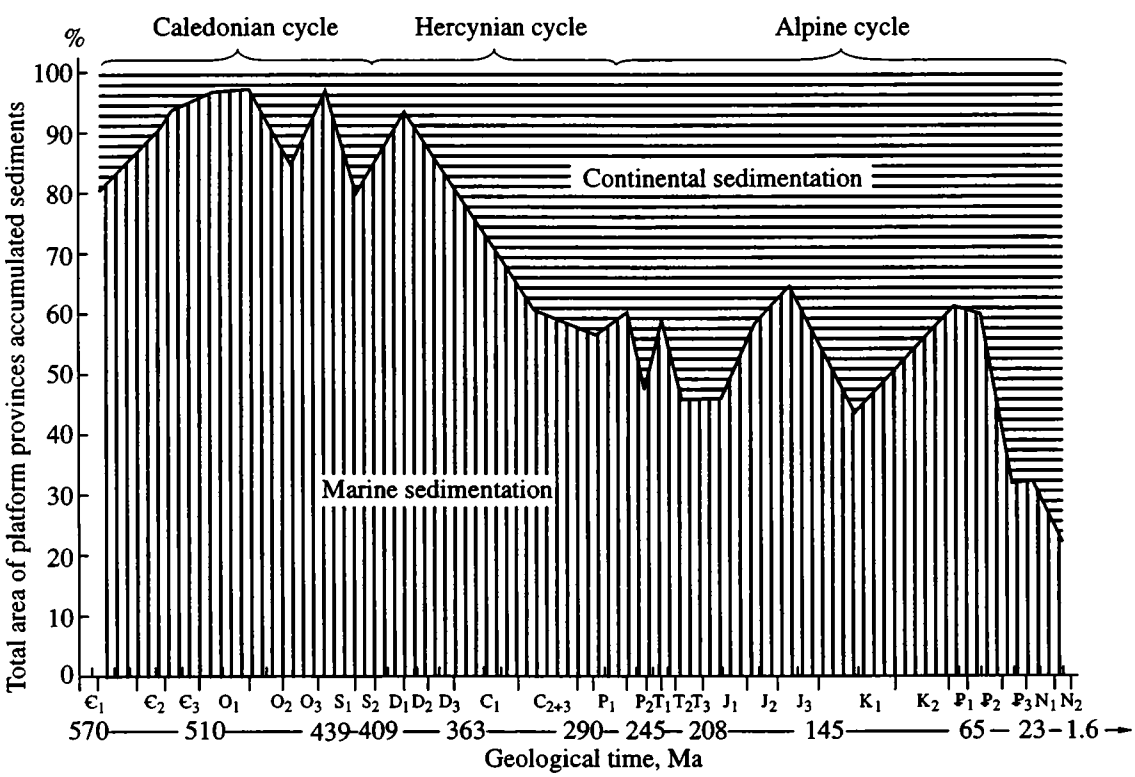

Fig. 10. Areal variations of marine and continental sedimentary provinces of continental platforms during the Phanerozoic.

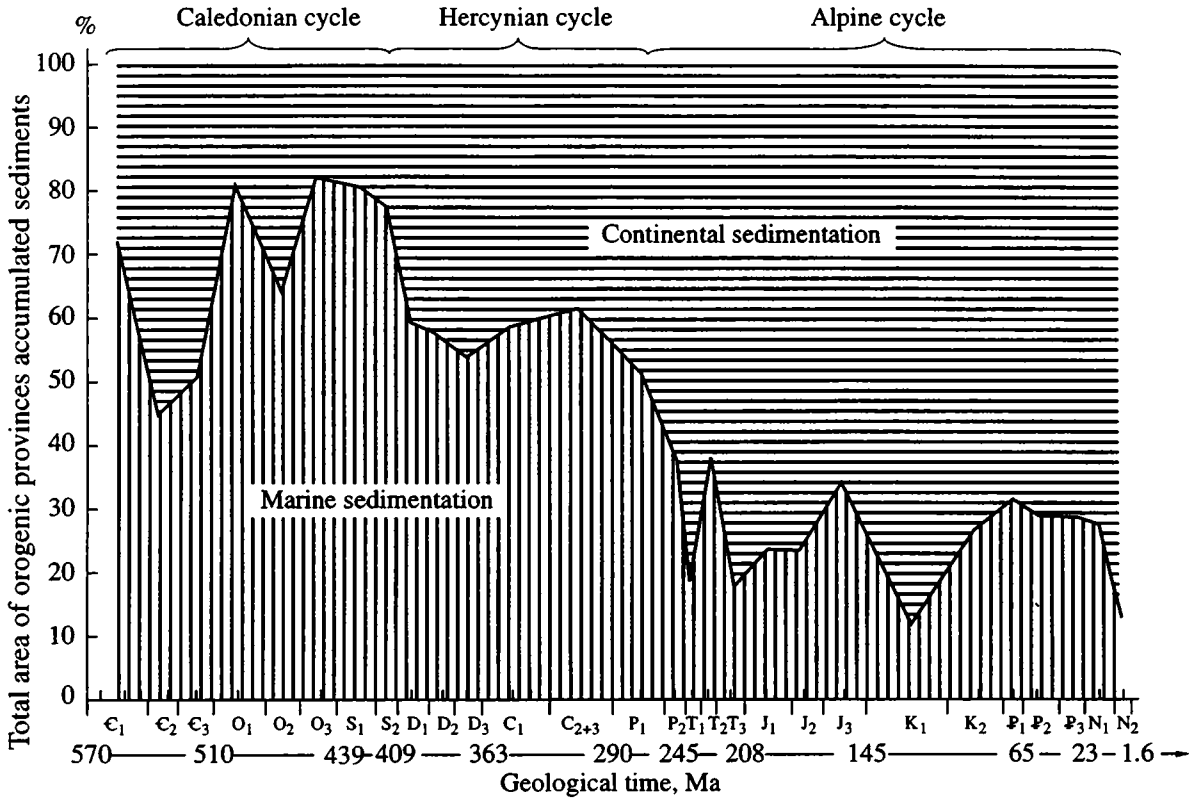

Fig. 11. Areal variations of marine and continental sedimentary provinces of continental orogens during the Phanerozoic.

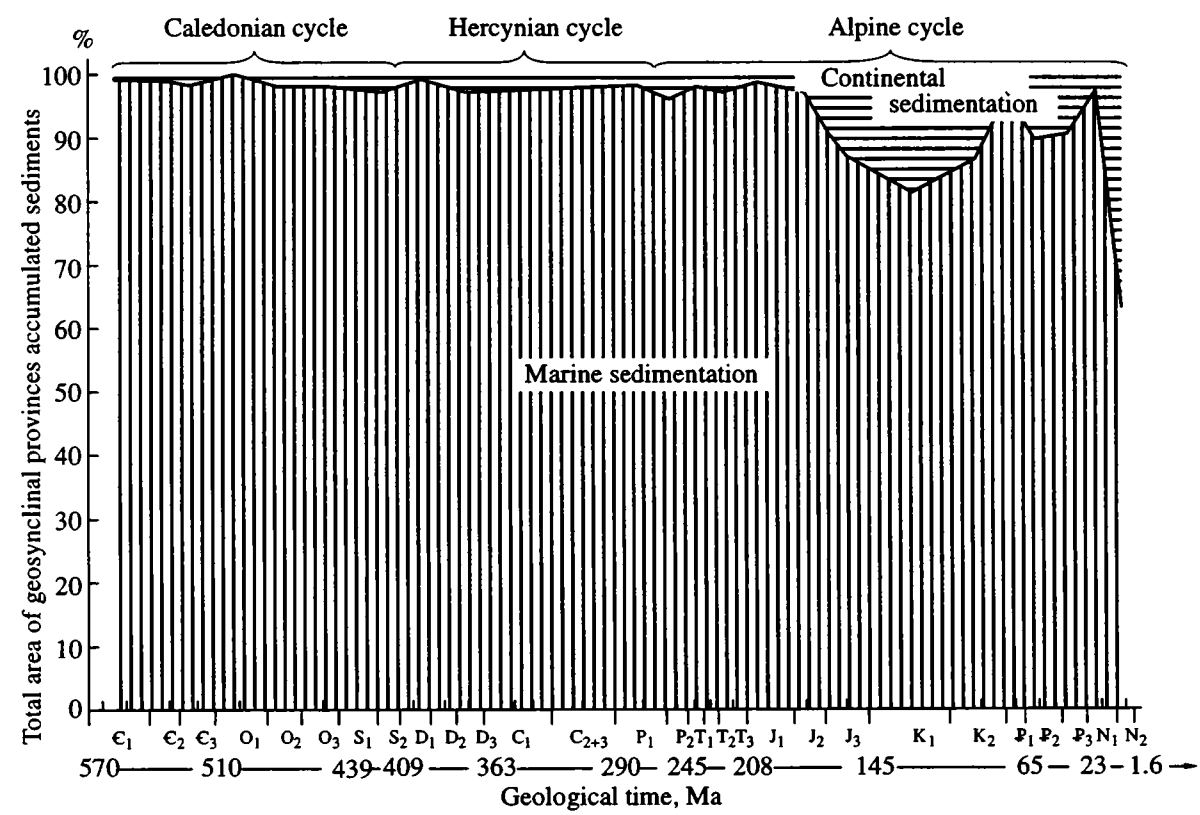

Fig. 12. Areal variations of marine and continental sedimentary provinces of geosynclinal zones during the Phanerozoic.

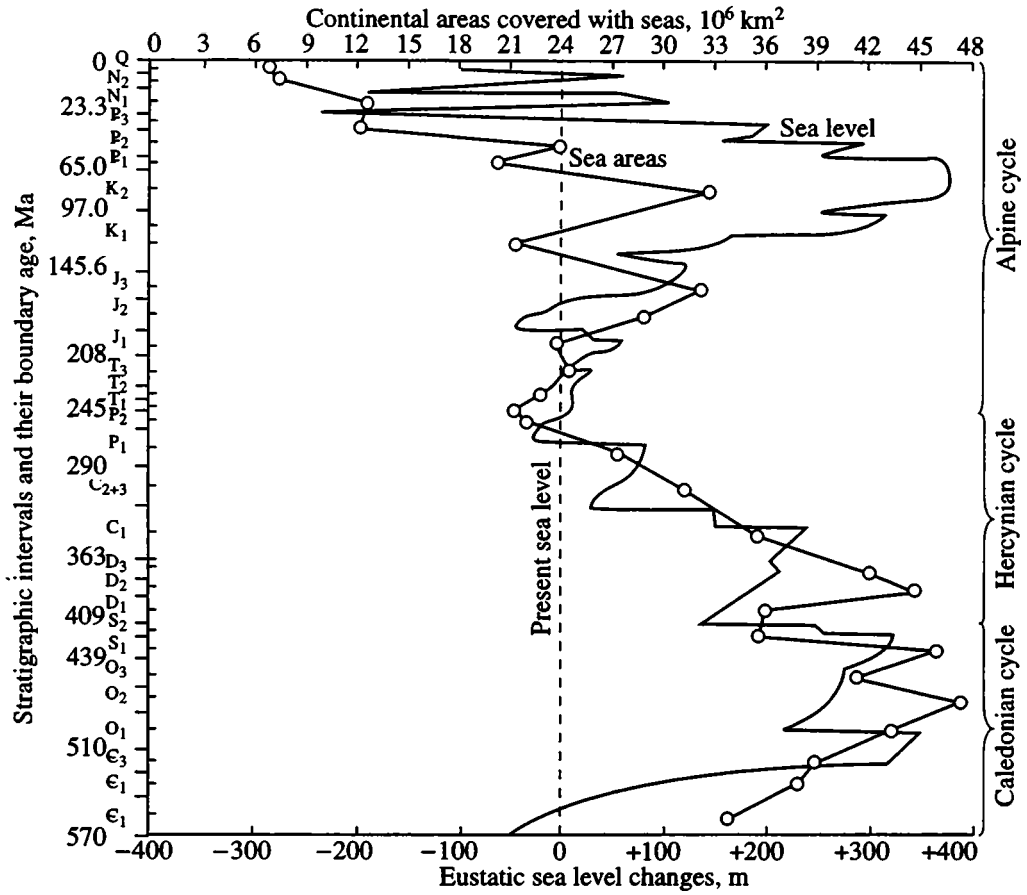

Fig. 13. Correlation between variation curves of sea transgressions and regressions (author's data), and sea level changes during the Phanerozoic (after Vail *et al.*, 1977).

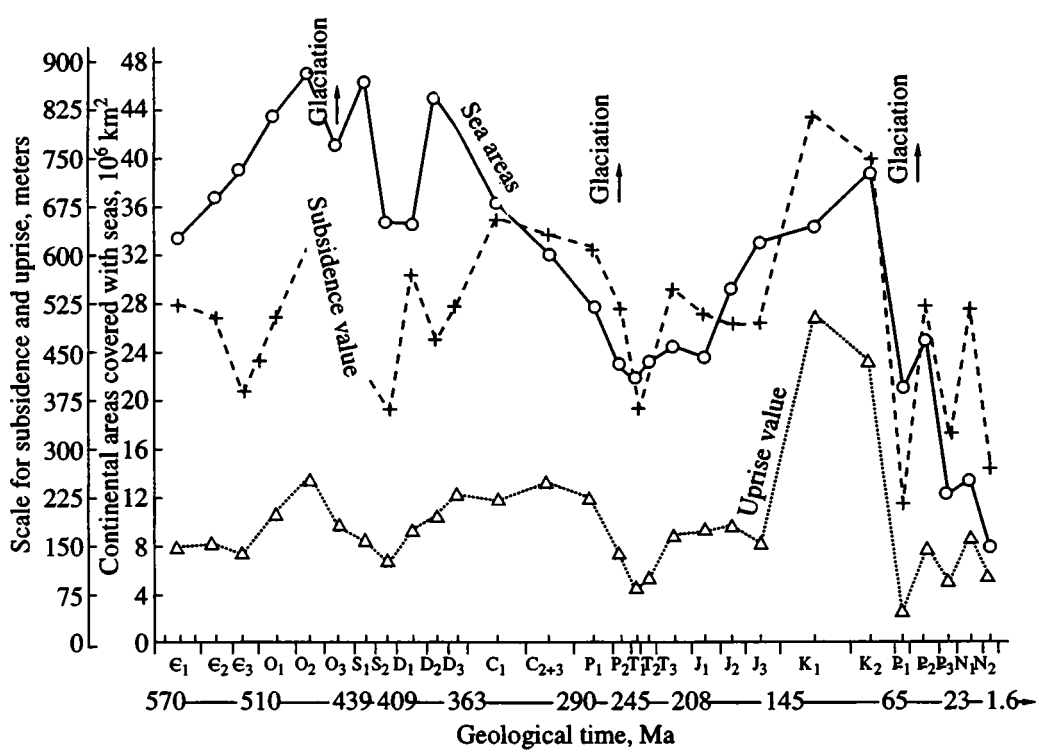
Scale for subsidence and uprise, meters
900 825 750 675 600 525 450 375 300 225 150 75 0
Continental areas covered with seas, 10^6 km^2
48 44 40 36 32 28 24 20 16 12 8 4 0
Glaciation
Sea areas
Glaciation
Glaciation
Subsidence value
Uprise value
ϵ_1 $\epsilon_2\epsilon_3$ O_1 O_2 O_3 S_1S_2 D_1 D_2D_3 C_1 C_{2+3} P_1 $P_2T_1T_2T_3$ J_1 J_2 J_3 K_1 K_2 P_1 $P_2P_3N_1N_2$
570 510 439 409 363 290 245 208 145 65 23 1.6
Geological time, Ma

Fig. 14. Variation curve of sea transgressions and regressions as compared to average continental subsidence and uprise (meters) at various epochs of the Phanerozoic (arrows mark epochs of worldwide glaciations).

In the future, we need to solve the problem in a new approach on the basis of quantitative studies.

Reviewer B.S. Sokolov

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New Data on Mezozoic–Cenozoic Stratigraphy of Eastern Kamchatka (the Valaginskii Ridge)

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Received April 15, 1993

Abstract – Composite stratigraphic sections of the upper Cretaceous–lower Eocene characterizing corresponding facial zones and those for overlying middle Eocene–Pliocene deposits in the central part of the Valaginskii Ridge are described. New data on the age of the Goluboi and Tal'nik members and the Vetlovaya Formation, based on radiolarians, benthic, and planktonic foraminifers, are presented. Previously unknown terrigenous facies of the middle Eocene are confirmed in the region. The Paleogene deposits of the Valaginskii Ridge are correlated with their reference Il'pinskii section of northeastern Kamchatka. Paleogeographical and geological events of the Vetlovian time are described.

INTRODUCTION

The Valaginskii Ridge, one of the eastern ridge systems, is an important element of the Kamchatka tectonic structure. It is impossible to speculate on the evolution of the eastern, much less the entire Kamchatka Peninsula without understanding the structure of the Valaginskii Ridge. Studies on its stratigraphy by V.I. Tikhonov, A.G. Tsikunov, B.I. Slyadnev, M.I. Goryaev, M.E. Boyarinova, and others revealed that siliceous volcano-terrigenous deposits of the upper Cretaceous (an upper part of which was later referred to the lower Paleocene) and Neogene terrigenous deposits are widespread throughout the region. During 1991 - 1992, we obtained new data concerning the composition and age interrelations between the stratigraphic units of this area.

The age relations and composition of the units were studied in the course of a large scale geological survey in the central part of the ridge, which includes the headwaters and watershed of the Valagina and Levaya rivers, and in Kitil'gina, Koyanova, and Konstantinovskaya river basins. M.K. Bakhteev, O.A. Morozov, S.R. Tikhomirova, and A.E. Shantser participated in field and laboratory investigations.

The age of the stratigraphic units was determined according to foraminifers, radiolarians, and mollusks. Foraminifers and radiolarians were found in rare interlayers of dense, sometimes recrystallized limestones and siliceous rocks. The study of carbonate rocks in thin sections showed that some lenslike interlayers contain abundant, well preserved foraminiferal tests and rare, calcite-substituted radiolarian skeletons. We used concentrated (45%) hydrofluoric acid (HF) to separate them. The samples with microfauna remnants visible in thin sections were splintered into pieces measuring 2 - 3 cm, which were then placed into plastic containers with acid for 10 - 15 minutes. After removing the acid, the residual was washed properly with water. As the amount of residual after a single leaching was insufficient, the process was repeated several times. The same portion of acid was not used more than three times.

Radiolarian skeletons were extracted from siliceous rocks by pickling them in a weak hydrofluoric acid (3 - 5%). The samples, splintered into the pieces measuring 3 - 5 cm, were kept in an acid for 10 - 12 hours. The residual was then washed properly with water.

V.N. Ben'yamovskii studied the foraminifers, N.Yu. Bragin and D.I. Vitukhin, the radiolarians, and V.N. Sinel'nikova, the mollusks.

RESULTS

Volcano-terrigenous deposits of Late Cretaceous, Paleocene, early-middle Eocene, Miocene, and Pliocene age compose the geological structure of the region. The Paleocene deposits were subdivided into subseries and stages according to Naidin and Ben'yamovskii (1988). The upper Cretaceous–lower Eocene deposits belong to different (northwestern and southeastern) facial zones divided by the South Valaginskii overthrust, which dips northwest at an angle of 40°. The tectonically formed sedimentary melange zone, up to 3.5 km wide (Fig. 1), is related to the overthrust. Outside the studied area, it includes tectonic lenses of serpentinite melange (Zinkevich *et al.*, 1992), indicating the deep roots of the zone.

UPPER CRETACEOUS–LOWER EOCENE

The Cretaceous–lower Eocene deposits, which have a total thickness of 2700 m, are present in the north-

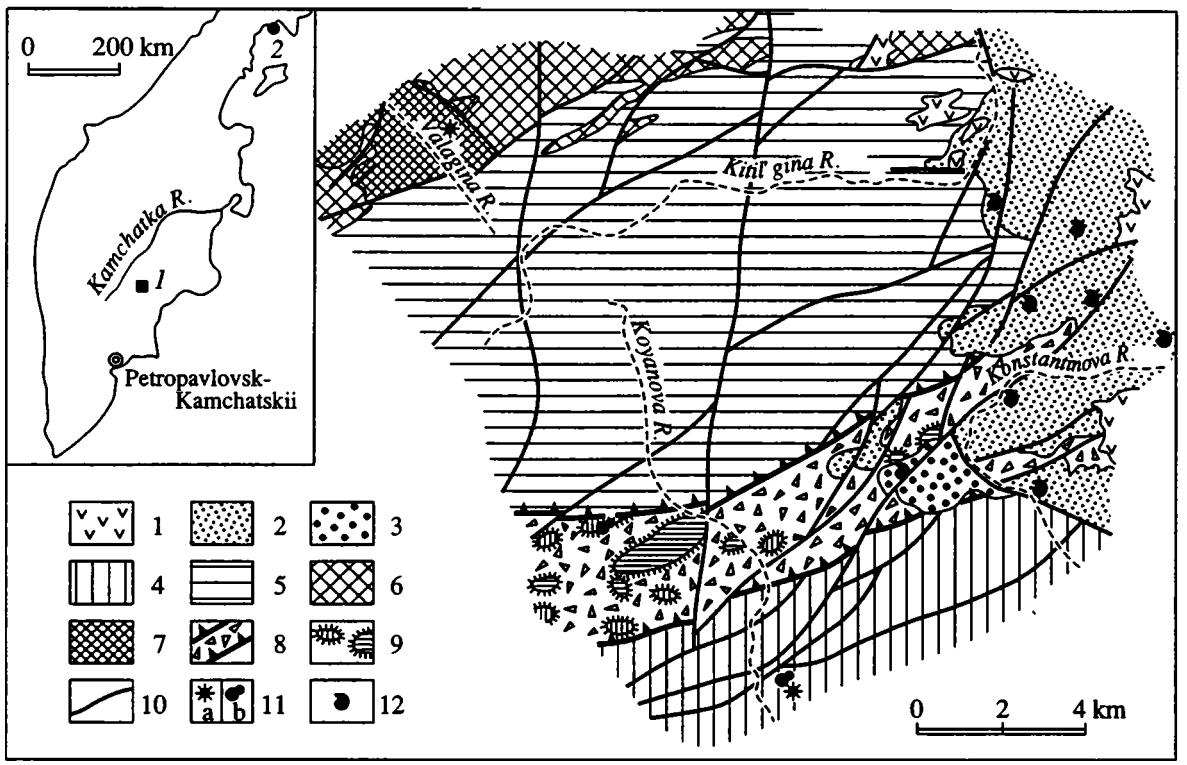

Fig. 1. Geological Scheme of the Central Valaginskii Ridge.

1 – upper Miocene–Pliocene, Shchapina Formation; 2 – lower-middle Miocene, Osipovskaya Member; 3 – middle Eocene; 4 - 5 – Paleocene–lower Eocene: 4 – Vetlovaya Formation, 5 – Tal'nik Member; 6 – Paleocene, Kitil'gina Formation; 7 – Upper Cretaceous–lower Paleocene (?), Goluboi Member; 8 – zone of the South Valaginskii overthrust and tectonic sedimentary melange; 9 – large blocks of the Vetlovaya Formation and Tal'nik Member within the melange; 10 – normal, reverse, and strik-slip faults; 11 - 12 – finds of: 11 – microfauna (a – radiolarian, b – planktonic and benthic foraminifers), 12 – malacofauna.

Inset map shows location of (*1*) the area considered and (*2*) Il'pinskii reference section.

western facial zone comprising the upper courses of the Valagina, Kitil'gina, Koyanova, and Konstantinova rivers, and the axial part of the Valaginskii Ridge. The Goluboi Member of tuffo-siliceous rocks with a total thickness exceeding 350 m occurs at the section base. The member consists of greenish grey and grey thin-bedded tuffo-siliceous and clayey siliceous rocks with interlayers of grey and black cherts, rare basaltic tuffs, and siliceous tuffites. The amount of the latter increases in the upper part of the member. The sediments show varying degrees of propylitization and carbonatization.

This member and the overlying Kitil'gina Formation form the Valagina Group. Its age was considered to be Senonian–pre-Maestrichtian according to rare prismatic inoceramyds and some radiometric datings (Shapiro *et al.*, 1987).

Siliceous interlayers in the middle part of the section on the right bank of the Valagina River contain the radiolarian species *Cornutella* cf. *californica* (Campbell et Clark), *Cromyodruppa* sp., *Cromyosphaera* sp., *Lithomitra* sp., *Stichomitra* cf. *livermorensis Campbell* et Clark, and *Stylosphaera* cf. *pusilla* Campbell et Clark of late Campanian–Maestrichtian and, probably, Paleocene age. Outside the studied area, in the northern part of the Valaginskii Ridge, the cherts in the lower section, partly corresponding to the Goluboi Member (Tsukanov, 1991), include typical Cenozoic radiolarian species of the *Ommatogramma* genus together with *Stichomitra livermorensis* Foreman, *Amphipyndax stocki* (Campbell et Clark), *Dictyomitra multicostata* Zittel gr., *Orbiculiforma* sp., *Stichomitra* sp., *Spongurus* sp., *Lithomitra* sp., and abundant species of the *Prunobrachium* genus. The latter includes forms similar to *Prunobrachium* cf. *longum* of Campanian–Paleocene age. The mentioned data allow us to estimate the Goluboi Member age as the late Campanian–early Paleocene.

The Kitil'gina Formation with abundant tuffs is 1300 m thick and conformably overlies the described deposits. It consists of two parts. The lower one, which constitutes 70% of the section, is mainly pyroclastic. It is represented by greenish-grey psammitic and vaguely laminated, basaltic or andesitic agglomerate tuffs with rare interlayers of cherry-grey and dark red andesites, basalts, their lava breccias, clastic lavas, and thin lenses of green cherts. The upper part is poorer in pyroclastic rocks. It consists of intercalating andesites and basalts with tuff interlayers.

In the section of the northwestern zone, the uppermost Tal'nik Member (more than 1000 - 1100 m thick) conformably overlies the Kitil'gina Formation. Mainly

sandy-clayey, it is ithologically subdivided into four units.

The coarse laminated structure is characteristic of the lower unit (170 - 240 m thick). The unit displays intercalation of grey, bluish-grey greywacke sandstones, thin laminated and massive gravelstones, siltstones and claystones, tuffaceous rocks, and tuffites, randomly bearing debris of basalts and limestones together with gravel sized quartz grains.

The second unit (400 m) is characterized by a coarse cyclic alternation of thick (up to 30 - 50 m) layers of black siltstone and interlayers (3 - 6 m thick) of grey to bluish-grey sandstones (feldspar-quartz greywacke) and sometimes tuffaceous sandstones.

The third unit (170 - 250 m) is represented by two- and, rarely, three-component flysch (with rhythmic units 0.5 - 1 m thick). The flysch displays alternations of grey feldspar-quartz greywacke sandstones with graded bedding, parallel to cross-bedded siltstones, and rare claystones. Sometimes rhythmic units are intercalated with relatively thick (1 - 1.3 m) layers of dark grey tuffaceous sandstones.

Dark grey, nearly black silty claystones, mudstones, and siltstones predominate the fourth unit (more than 190 m). The rocks include thin interlayers of subarkosic sandstones and limestone concretions.

The stratigraphic position of the described deposits is debatable. Under the name of the Vakhvina Group, some scientists place them below the Valagina Group (*Resheniya ...*, 1974, 1984; Zinkevich *et al.*, 1992). The others consider them to be part of the Vaskuchevskaya Formation overlying the Kitil'gina Formation (Petrov *et al.*, 1966). And finally, some think that the Vakhvina Group includes two terrigenous units, one underlying and the other overlying the Valagina Group (Seliverstov and Grechin, 1979).

We believe that in the studied area the deposits form an indivisible unit overlying the Kitil'gina Formation. This is indicated by the fact that its base and basal debris members are traceable throughout the area between the middle course of the Kitil'gina River and the Levaya River basin. The uniform structure of this stratigraphic unit and persistent lithologic characteristics also support our interpretation. There are stratigraphic contacts of the studied deposits with the siliceous rocks of the Goluboi Member, which would be expected if they included older terrigenous rocks underlying the siliceous volcanogenic sediments of the Valagina Group. According to composition and structure, the described deposits correspond to the Tal'nik Member distinguished by M.E. Boyarinova in the basin of the Bol'shoi Tal'nik Creek.

In 1992, in sample (2419/4) from the Tal'nik Member located west of the studied region, Boyarinova determined the following planktonic foraminiferal assemblage: *Acarinina acarinata* Subb.– *A. subsphaerica* Subb., *Globigerina nana* Chal., *G. velascoensis* Cushman, *G. bacuana* Chal., *Globorotalia reissi* Loebl. et Tapp.–*G. perclara* Loebl. et Tapp. The assemblage suggests the late Paleocene–early Eocene age of the sediments (Table 1).

In the central part of the Valaginskii Ridge, the Tal'nik Member composes a united section with the Goluboi Member and the Kitil'gina Formation. The section has no angular or visible stratigraphic unconformities. The Tal'nik Member lies significantly higher than layers with the late Campanian–Maestrichtian radiolarians and probably above the Paleocene sediments of the Goluboi Member. All this allows us to date the Tal'nik Member as the Paleocene–early Eocene in age.

The stratigraphic position of the Kitil'gina Formation between beds with upper Cretaceous–Paleocene (?) and upper Paleocene–lower Eocene fauna indicate that it corresponds to a part of the lower Paleocene.

Sediments of the Vetlovaya Formation are exposed in the southeastern facial zone covering the territory of the middle course of the Koyanova and Konstantinova rivers. The lower boundary of the formation and underlying sediments are not exposed. From the Tal'nik Member, the formation is separated by a tectonic melange zone. The formation structure is clearly visible in the continuous outcrops along the canyonlike slopes of the main and tributary valleys of the Koyanova and Konstantinova rivers (in their middle courses). This mainly carbonate-siliceous-volcanogenic formation is subdivided into three units.

The lower (tuffo-sedimentary) unit is represented by greenish-grey, faintly laminated, unequigranular tuffaceous sandstones with lenslike interlayers of calcareous gravelstones, tuffaceous claystones, tuffo-siliceous rocks, and silty limestones. Its apparent thickness is 550 m.

A distinct thin and parallel bedding distinguishes the middle (siliceous-carbonate) unit (400 - 450 m) from the underlying and overlying deposits. The unit includes grey, greenish-grey carbonate-siliceous, clayey-siliceous, and siliceous-carbonate rocks containing a tuffaceous admixture and interlayers of organic detrital limestones, cherts, volcaniclastic sandstones, and gravelstones. Abundant planktonic foraminifers were found in the layer of pelitomorphic limestones (sample 2425) near the source of one of the right tributaries of the Koyanova River. Species of *Acarinina mckannai* (White) (*A. subsphaerica* Subb.), *A. soldadoensis* (Bronn.), *A. acarinata* Subb., *Globorotalia aequa* Cushm. et Renz., *G. pseudomenardii* Bolli, *G.* cf. *velascoensis* (Cushman), and *Globigerina nana* Chal. were determined in this assemblage.

All the species except *Globorotalia pseudomenardii* are known from the upper Paleocene (Thanetian) deposits. The last form has a short stratigraphic range and is typical of the lower Thanetian *G. pseudomenardii* Zone. This indicates that the second unit is early Thanetian in age (Table 1). A nearly identical assemblage was determined in the upper subformation of the South-Il'pinskii Formation of the Paleogene section

Table 1. Stratigraphic position of studied samples of the Tal'nik Member and the Vetlovaya Formation

Series	Subseries	Stage	International planktonic foraminiferal scale	Regional planktonic foraminferal scale	Stratigraphic position of the samples	
					Vetlovaya Formation	Tal'nik Member
Eocene	Lower	Ypresian	*Acarinina pentacamerata*		2726/2	
			Globorotalia aragonensis	*Pseudohastigerina wilcoxensis*		
			G. formosa			
			G. subbotinae	*Globorotalia wilcoxensis*		2419/4
Paleocene	Upper	Thanetian	*G. velascoensis*			
			G. pseudomenardii	*Globigerina nana*	2425	
		Zealandian	*G. pusilla*			
			G. angulata			
	Lower	Montian	*Acarinina uncinata*			
		Danian	*Globorotalia trinidadensis*			
			G. pseudobulloides			
			Globigerina eugubina			

located in the Il'pinskii Peninsula (Ben'yamovskii *et al.*, 1992) (Table 2).

The radiolarian species *Theocotyle* aff. *nigrinae* Riedel et Sanfilippo typical of the lower-middle Eocene of the tropical Pacific provinces was determined in the same sample. The Eocene–Oligocene species *Stylosphaera minor* Campbell et Clark occurs less frequently.

The problem of discordant age estimations from data on radiolarians and foraminifers needs further investigation.

Numerous representatives of the *Globorotalia* genus (more than 30% of the total test number) indicate the warm water regime of the basin. Such forms as *Globorotalia pseudomenardii*, *G. aequa*, and *G. velascoensis* are typical tropical forms. In the Il'pinskii reference section, only *G. aequa* (single tests) was found. This shows that lower Thanetian sediments of the studied area (middle member of the Vetlovaya Formation) accumulated within the tropical–subtropical zone of the ocean. Planktonic forms sharply predominate (nearly 100%) in the foraminiferal paleocoenosis (assemblage). Such assemblages could only have existed in the pelagic part of the open ocean.

The upper (tuffaceous) member (more than 530 m) mainly consists of dark grey or greenish-grey massive, faintly laminated, silty to silty-psammitic tuffs and tuffites of andesite-basaltic composition with subordinate interlayers of siliceous rocks, ferruginous mudstones, limestones, flows of andesite-basalts, leucobasalts, subalkalic olivine basalts, and giant lentil-like concretions of micrograined limestones.

Similar to sediments of the Tal'nik Member, the rocks of the middle and upper units of the Vetlovaya Formation form large (hundreds of meters in diameter) blocks included in the tectonic sedimentary melange separating facial zones. These blocks show well preserved peculiarities of the bedding, composition, and inner fault-folded structure of the original stratigraphic sections. In the area of the upper stream of the Levaya River, planktonic foraminifers were found in the limestone interlayer of the upper unit of the Vetlovaya Formation observed in such a block (sample 2726/2). Their assemblage includes the following species: *Pseudohastigerina wilcoxensis* Cushm. et Pont., *Acarinina* cf. *pentacamerata* Subb., *A.* cf. *pseudotopilensis* Subb., *Globorotolloides suteri* Bolli–*Acarinina broedermanni*, *Globigerina* cf. *boweri* Bolli, *G. patagonica* Todd and Kniker, *G. pseudoeocaena* Subb. (*juv.*), and *G. inaequispira* Subb. (*juv.*). The characteristic benthic species *Bulimina aksuatica* Moroz. and *Globocassidulina globosa* Hantk. were also found.

Undoubtedly, the species composition of this assemblage allows us to correlate the upper unit with the Ypresian Stage of the lower Eocene. The assemblage possibly corresponds to the *Pseudohastigerina wilcoxensis* Zone of the Il'pinskii section (Fig. 2; Tables 1 and 2). The *Pseudohastigerina* datum marks the Eocene lower limit in northwestern Europe (Berggren, 1967). The described benthic foraminifers do not

Table 2. Correlation of the planktonic foraminiferal assemblages

Genus	Species	Zone	Il'pinskii section	Vetlovaya Formation	Tal'nik Member
Globigerina	*Globigerina nana*	*Globigerina nana* (P_1t)	+	+	+
	G. velascoensis		+		+
	G. incisa		+		
	G. triloculinoides		+		
	G. bacuana		+		+
	G. compressaeformis		+		
	G. angipora		+		
Globorotalia	*Globorotalia aequa*		+	+	
	Gl. pseudomenardii			+	
	Gl. cf. *velascoensis*			+	
Acarinina	*Acarinina acarinata*		+	+	+
	A. mckannai (= *A. subsphaerica*)			+	+
	A. soldadoensis			+	
Globorotalia	*Globorotalia reissi–G. perclara*	*Globorotalia wilcoxensis* (P_2ip_1)	+		+
Globigerina	*Globigerina inaequispira*	*Pseudohastigerina wilcoxensis* ($P_2ip_1^2$)	+	+	
	G. ps. *pseudoeocaena*		+	+	
	G. gr. *turgida*		+		
	G. patagonica		+	+	
	G. linaperta		+		
	G. posttriloculinoides		+		
	G. aff. *incretacea*		+		
	G. aff. *eocaena*		+		
Pseudohastigerina	*Pseudohastigerina micra*		+		
	P. wilcoxensis		+	+	
Globorotaloides	*Globorotaloides suteri*		+	+	

dispute this supposition, because their appearance in the Il'pinskii section is at the level of the Ypresian deposits.

Paleoecological analysis of the fossil foraminifers reveals a significant reorganization of the paleoecosystem during the late Ypresian time. A warm tropical-subtropical water mass with abundant species of the *Acarinina* and *Globorotalia* genera was replaced by a cooler (subtropical-temperate) one inhabited by species of the *Pseudohastigerina, Globigerina,* and *Globorotalloides* genera typical of the temperate-warm seas of middle latitudes. Species of the *Pseudohastigerina* genus are numerous in the high latitude assemblages (Toumarkline and Luterbacher, 1985).

Those scientists that study the described sediments of the Koyanova and Konstantinovskaya river basins usually refer them to the Vetlovaya Formation. Boyarinova considered the Vetlovaya Formation section situated to the west of the studied area on the left bank of the Levaya River to be its hypostratotype. Kovel' identified *Globigerina quadritriloculinoides* Chal., *G. incisa* Hill., *G. pseudoeocaena* Subb., *G. inaequispira* Subb., *Acarinina acarinata* Subb., *A. soldadoensis* (Bronn.), *A. subsphaera* Subb., *Globorotalia wilcoxensis* Cushm. et Pont., and *G. reussi* Loebl. et

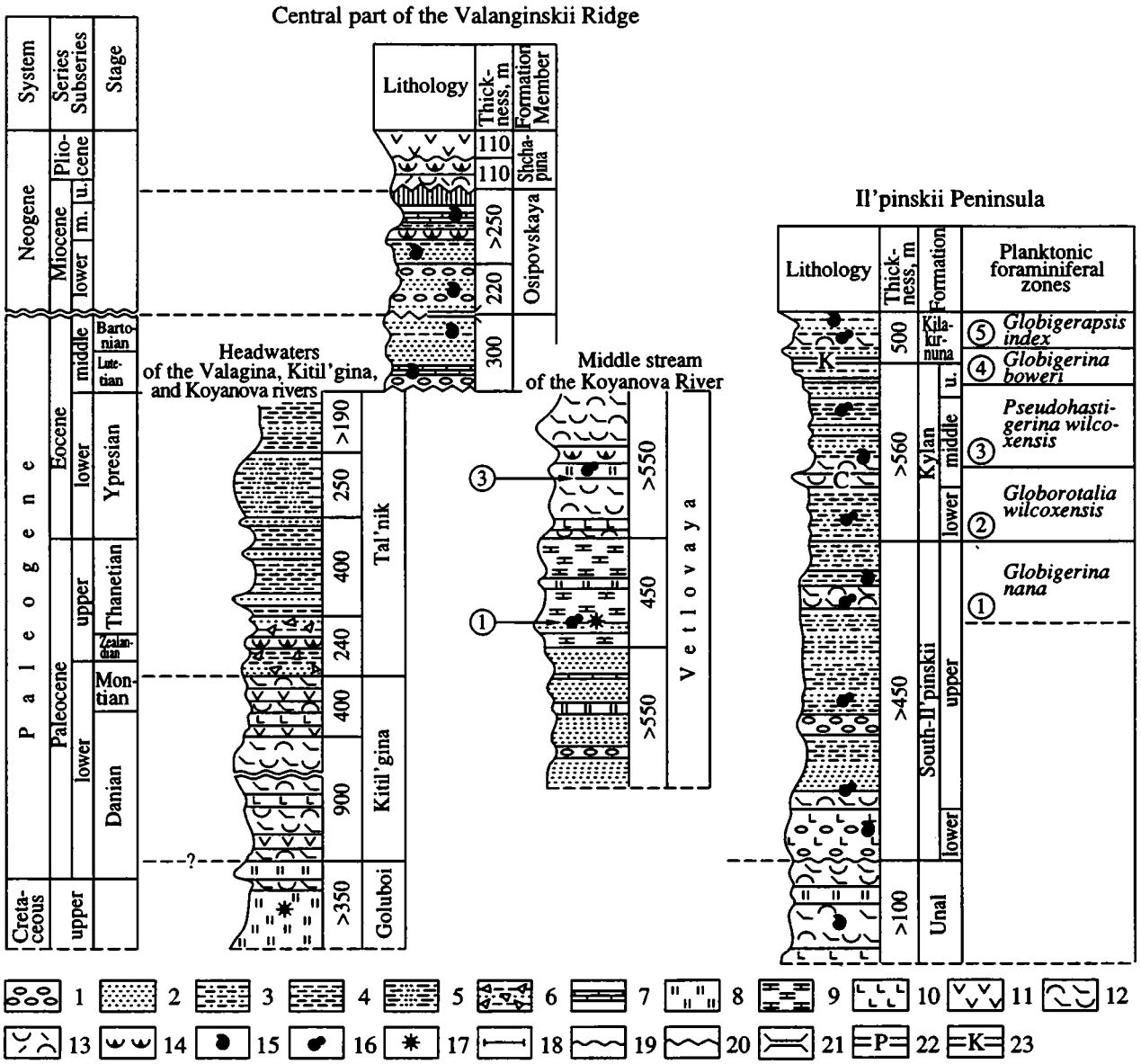

Fig. 2. Correlation of the stratigraphic sections of the central Valaginskii Ridge and Il'pinskii Peninsula.

1 – conglomerates, tuffaceous conglomerates, gravelstones; 2 – sandstones, tuffaceous sandstones; 3 – siltstones; 4 – clays, claystones, tuffaceous claystones; 5 – sand-silt-claystone flysch; 6 – rubble detrital sandstones and claystones; 7 – limestones; 8 – siliceous, clayey-siliceous, tufaceous-siliceous rocks; 9 – siliceous-calcareous, calcareous-siliceous rocks; 10 – basalts; 11 – andesites and basaltic andesites; 12 – tuffs of basic and intermediate composition; 13 – acid tuffs; 14 – tuffites; 15 – mollusks; 16 – foraminifers; 17 – radiolarians; 18 - 21 – type of boundaries between units (18 – conformable, 19 – erosional, 20 – angular unconformities, and 21 – tectonic contacts); 22 - 23 – marker horizons (22 – lilac tuffs and 23 – tuffaceous Kilakirnuna Sandstone).

Tapp. in sediments of the upper Vetlovaya Formation in the Kavycha River valley within the territory adjacent to the hypostratotype and dated the sediments at the early Eocene.

One of the samples from M.E. Boyarinova's collection (2542/14) obtained from the Vetlovaya Formation sediments in the same region contains planktonic foraminifers of the *Globigerina triloculinoides* Plumm. group and poorly preserved specimens of *Globigerina varianta* Subb., *Globorotalia* cf. *trinidadensis* Bolli, and *Globoconusa* cf. *daubjergensis* (Bronn.), characteristic of the Danian Stage of the Paleocene (the *Globorotalia trinidadensis–Globoconusa daubjergensis* Zone according to Ben'yamovskii *et al.*, (1992). However, the poor preservation and rare occurrences of the foraminifers throw doubt on the reliability of the determinations

Paleontological data suggest the Paleocene–early Eocene age of the Vetlovaya Formation. It probably corresponds to the Tal'nik Member together with the upper part of the Kitil'gina Formation of the adjacent northeastern facial zone.

The younger middle Eocene, lower-middle Miocene, and upper Miocene–Pliocene deposits are identical in both facial zones.

MIDDLE EOCENE

As mentioned above, the facial zones are separated by the melange zone. Inclusions in the tectonizated claystonelike matrix consist of different rocks of the Tal'nik Member and Vetlovaya Formation. The melange is unconformably overlain ("sealed") by the tuffaceous claystone and sandstone beds, which are mainly represented by greenish-grey graywacke sandstones and silty tuffaceous claystones. The basal, lenslike bed consists of dark grey small-pebble conglomerates, gravelstones, and sandstones. The latter contain abundant and variably preserved mollusks sometimes forming coquinites. Among them, *Miltha snatolensis* Sin., *Modiolus flabellatus utcholokensis* Slod., *Tellina snatolensis* Sin., *Macoma* sp., and *Mya* sp. were identified. These mollusks are index fossils of the Snatol Horizon in western Kamchatka and indicate the middle Eocene age of the beds. Within the Valaginskii Ridge, the middle Eocene terrigenous deposits were recognized for the first time. They are over 300 m thick.

LOWER-MIDDLE MIOCENE

The lower-middle Miocene deposits are represented by the Osipovskaya Member of the Kornilovskaya Group, which can be traced from the Osipovskaya River to the headwaters of the Konstantinova River. The member unconformably overlies sediments of the Kitil'gina Formation, the Tal'nik Member, and the Vetlovaya Formation and has tectonic contacts with the middle Eocene deposits. This mainly terrigenous member is subdivided into two units.

The lower (basal) unit (up to 220 m) is represented by yellowish-grey, brownish-grey, and bluish-grey inequigranular greywacke sandstones with numerous lenses and interlayers of brownish-grey gravellestones and pebbly and cobble-pebbly, inequiclastic and pudding conglomerates. The upper unit (more than 250 m) is composed of dark grey and tobacco-grey tuffaceous sandstones intercalating with siltstones and claystones, containing tuffite interlayers, lenses and concretions of sandy and silty limestones, calcareous sandstones, and siltstones. The rocks are full of "floating" gravel and pebbles and small (a few centimeters thick) lenses of gravelly sandstones, gravelstones, and small-pebble conglomerates.

Lenses and concretions of the calcareous rocks contain abundant remnants of *Cardium etheringtoni* Kog., *C. meekianum* Gabb., *Clinocardium nuttallii* (Conr.), *Macoma chejsliensis* Sin., *M.* cf. *optiva* Yok., *Macoma* aff. *orbiculata* Scarlato, *Mya cuneiformis* Bohm., *M. profundior* Grant et Gale, *Spisula equilateralis* Clark, *Tellina aragonica* Dall, and *Thiasira disjuncta* Gabb. Among them, *Mya cuneiformis, Spisula equilateralis*, and *Thiasira disjuncta* are characteristic of the Kuluven and Il'inskaya formations of the lower Miocene–initial middle Miocene in western Kamchatka. All the other forms occur in the sections of the middle Miocene Kakertskaya Formation. The age of the Osipovskaya Member is the early-middle Miocene.

UPPER MIOCENE–PLIOCENE

The end of the pre-Quaternary section is marked by the Shchapina Formation, which overlies the Osipovskaya Member with an angular unconformity. The basal unit (110 m) is represented by rhythmically alternating tuffites of hybrid composition, containing psammitic silty dacite tuffs and sandy clays. Tuffites contain rare and small scattered pebbles, gravel, and coalified detritus. The sediments of the upper unit (more than 130 m), represented by several generations of andesite and leucocratic basaltic andesite flows, fill the pre-volcanic relief depressions and occur nearly horizontal on the gently sloping sedimentary-tuffaceous deposits of the lower member and sandstones of the Osipovskaya Member. Previous studies (Volynets *et al.*, 1990) considered the age of the Shchapina Formation to be the late Miocene–Pliocene.

CONCLUSION

(1) The Cretaceous and Paleogene deposits of the northwestern structural-facial zone form a stratigraphically continuous section in which angular or structural unconformities are absent.

(2) The Paleocene–lower Eocene deposits of the upper part of the Valagina Group together with the overlying Tal'nik Member of the northwestern facial zone correspond to the sediments of the Vetlovaya Formation of the southeastern zone.

(3) The Vetlovaya Formation composes geological bodies with specific structures, rock sequences, and biostratigraphic characteristics, which can be distinguished by mapping.

(4) The Vetlovaya sedimentation stage corresponds to two large eustatic transgressions: the Thanetian–early Ypresian and the late Ypresian. The first one is characterized by a climatic optimum known not only in the northern Pacific, but in other parts of the northern hemisphere as well (Chelebaeva and Brattseva, 1986).

(5) The first structural reorganization within the Valaginskii Ridge since the Campanian took place at the early-middle Eocene boundary or at the beginning of the middle Eocene.

The authors thank V.S. Sverdlov for his assistance in field and laboratory studies.

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Eocene Deposits on the Northeastern Slope of the Dnieper–Donets Depression: Zonal Stratigraphy and Cyclic Analysis

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Received September 20, 1993

Abstract – In the Eocene, the northern slope of the Dnieper–Donets depression was an area of transitional carbonate to terrigenous-siliceous sedimentation. The information on foraminifers, radiolarians, diatoms, and silicoflagellates was used to redefine the age of the Kiev and the lowest unit of the Khar'kov (Derizovka) Group and to refer them to the upper Lutetian and the Bartonian. Both formations characterize two complete transgressive-regressive cycles beginning and terminating with sand deposits. The richest assemblage of microorganisms is confined to the lower beds of the Kiev Formation where the sands grade into clays. These beds contain nannoplankton, benthic foraminifer radiolarian and diatom assemblages of open-sea environments. All of the planktonic groups have zonal species that date the rocks. In the upper part of the Kiev Formation, the diatom assemblage is impoverished. The transgressive sequence of the second cycle contains radiolarians, diatoms, and silicoflagellates in transition from sands to silts of the Khar'kov Group. The radiolarian assemblage is poor, whereas the diatom community is supplemented by numerous neritic species. All of the three assemblages, however, contain open-sea index species useful for zonal subdivision.

INTRODUCTION

Two provinces are distinguished by the type of sections and the composition of microfossils of the Eocene sea basin that covered the southern part of the Russian platform and the surrounding areas. The section of carbonate type, which has diverse benthic and planktonic foraminifers and nannoplankton, is typical of the southern province. The northern province is characterized by mainly noncalcareous sandy-clayey sediments with sparse benthic foraminifers and nannoplankton and abundant assemblages of radiolarians, diatoms, and silicoflagellates.

The Dnieper–Donets depression is a classical area containing Paleogene deposits of the northern type. This Paleogene sequence has a single interval well characterized by microfossils, i.e., the Kiev Formation with diverse foraminifers and nannoplankton and the overlying lower beds of the Khar'kov Formation with abundant diatoms, silicoflagellates, and a smaller amount of poorly preserved radiolarians. The interval could have been a reliable reference unit for the Paleogene biostratigraphy of the Dnieper–Donets depression, but this was not realized for two reasons. First, the exact stratigraphic position of the Kiev Formation is still problematic. Second, the stratification of the overlying sequence with diatoms and radiolarians is determined by the local zonal scales based on siliceous organisms, and the age of these subdivisions is derived, eventually, from the age of the Kiev Formation. We will discuss this problem in detail.

LITHOSTRATIGRAPHY

The stratotype sections of the second half of the Paleogene are located in the Dnieper area near Kiev, where the Kiev Formation and the Khar'kov Group were recognized. The Kiev Formation, formerly known as the local "Kiev Stage" (Sokolov, 1893), includes three distinct beds, namely, the lower one of phosphoritic sand, the middle composed of the "Kiev marl," and the upper noncarbonate clay, the so-called "Naglinok." Initially, the late Eocene age of the Kiev Formation was determined by mollusks, but when foraminifers and nannoplankton were found in the Kiev marl, the sequence was referred to the middle Eocene.

The "Khar'kov Stage" was described as a unit of glauconite sands and noncarbonate clays and compared (Sokolov, 1893) to the Lattorfian glauconite sands by lithology and mollusks, i.e., it was dated as the Oligocene. Later, by indirect correlations, these deposits were assigned to the level of the Mandrikovka limestones, observed near Dnepropetrovsk and described as containing nannoplankton of the upper Eocene (Makarenko *et al.*, 1987).

In Russia in the 1950 - 1960s, Semenov's lithostratigraphic scale (1965) was used to subdivide these same deposits. The upper part of the Paleogene section was divided into the Sergeevka, Tishki, and Kas'yanovka formations. Their stratotype sections are located in the middle and lower reaches of the Severnyi Donets River. The Sergeevka and Tishki formations correspond to the Kiev Formation in the Ukraine, and the Kas'yanovka Formation, to the Khar'kov Forma-

tion. This correlation was rather confused, as it united the upper part of the Kiev Group with the Khar'kov Group.

The mapping of the territory has revealed that the Kiev Formation grades into clayey-siliceous deposits on the northeastern and southwestern slopes of the Dnieper–Donets depression, thus causing ambiguity in differentiation between the Kiev and Khar'kov beds. As a consequence, the layers with siliceous interbeds were incorporated into the upper Kiev Subformation (Table 1), and the Khar'kov Formation was discarded (Zosimovich *et al.*,1963). At a later date, the upper layer of the Kiev Formation, Naglinok, was assigned to the Khar'kov Group, which thus acquired two units, i.e., the lower Obukhovo Formation with all siliceous beds and the upper sandy Mezhigorka Formation with scarce organic remains (Makarenko *et al.*, 1987).

PREVIOUS BIOSTRATIGRAPHIC DATA

The foraminifers of the Kiev Stage (otherwise Horizon, or Formation) in the Dnieper–Donets depression were under study for more than half a century. Kaptarenko-Chernousova (1951, 1956) generalized the results of earlier research and described 120 species of foraminifers. Fursenko and Fursenko (1961) characterized 69 additional species in a report on foraminifers from the Kiev deposits in Belarus' and Lithuania. In both works, the deposits were dated as the late Eocene according to lithological correlations.

The foraminiferal assemblages of the Kiev Stage were stratigraphically analyzed in the publications by Nikitina (1963), Kraeva (1974), and Kraeva and Lyul'eva (1977). According to the scale adopted at that time in the Ukraine, Kraeva divided the Kiev Formation into two subformations. She compared the assemblages of foraminifers of the lower Kiev Subformation with the zones of *Hantkenina alabamensis* and *Globigerina turcmenica* of the Keresta and Kuma horizons in the Crimea–Caucasus scale. The upper member was regarded as an analog of the *Globigerapsis tropicalis* Zone of the Belaya Glina Horizon. Nikitina referred only the lower subformation to the Kiev Stage, whereas the upper one was referred to the Khar'kov Stage. In later papers on the Paleogene of the western and southern parts of the East European platform, the Kiev Formation was dated as the upper Lutetian–Bartonian (zones P12 - P14) according to foraminifers and nannoplankton (Makarenko *et al.*, 1987; Grigyalis *et al.*, 1988; Zosimovich, 1992).

Consequently, biostratigraphers studying the foraminifers were of different opinions concerning both the stratigraphic range of the Kiev Formation and the correlation of both formations with the horizons of the Crimea–Caucasus province (Table 1).

The radiolarians, recovered from the Kiev and Khar'kov formations in the northeastern Ukraine and adjacent areas of Russia, have been known since the end of the last century. Kaptarenko-Chernousova (1951) was the first to attribute the noncarbonate radiolarian clays, overlying the so-called Kiev marl, to the Kiev Stage rather than to the Khar'kov Stage.

Tochilina (1966) described a number of new radiolarian species and genera from these rocks and was the first to identify four horizons (beds) with different radiolarian assemblages, i.e., the Sumy, Kanev, Kiev, and Khar'kov horizons (Tochilina, 1975). She compared the radiolarian assemblage of the Kiev Formation with the Kuma assemblage in Turkmenistan, and the Khar'kov assemblage with the Belaya Glina assemblage from provinces of the lower reaches of the Don River and the eastern Caspian region. Gorbunov (1979) identified three radiolarian assemblages from the lower part of the Buchak Formation and from the middle and upper parts of the Kiev Formation. He dated the Buchak assemblage as the middle Eocene, and both Kiev assemblages as the late Eocene, thus following the lithostratigraphic correlation of the units. This is the reason why we refer to Kozlova when discussing the observation data (Kozlova and Gorbovets, 1966; Kozlova, 1990; and others). Her zonal radiolarian scale of the boreal zone (see the figure) incorporates the data of all previous studies on radiolarians from the Paleogene deposits of Russia and the adjoining areas. Though we were unable to determine in this work the zones of the boreal scale because of the lack of studied radiolarian taxa, the identified communities still more or less correlate with Kozlova's zonal assemblages.

The first diatoms were recovered from the stratotype section of the Khar'kov Group (Uspenskaya, 1935; 1940), which yielded three assemblages identified by Jousé (1949). Their stratigraphic range was not indicated, but the upper assemblage was noted to be richer in numerous species of oceanic plankton.

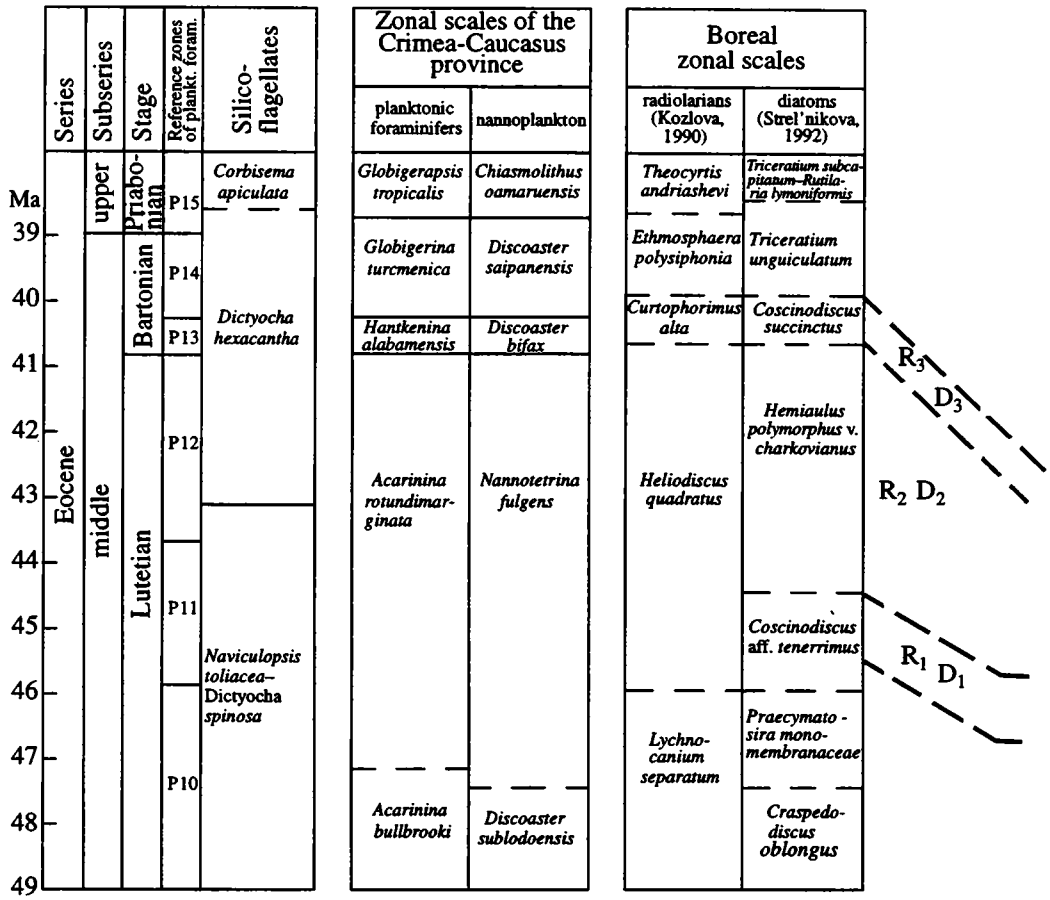
Ol'shtynskaya (1978) attributed the diatoms from the Tishki and Kas'yanovka formations to the lower and upper Kiev assemblages. The lower assemblage was characterized as impoverished and the upper one as having a high species diversity at the expense of a higher proportion of pennate diatoms and silicoflagellates. Both were dated as the late Eocene.

In the stratotype Staroverovka section along the Northern Donets River, Glezer assigned the upper Kiev Subformation assemblage to the *Paralia oamaruensis* Zone by correlating it with the oceanic diatom assemblages of Barbados Island and the Oamaru Group (New Zealand). She considered the unit's age first to be the early Oligocene (Glezer *et al.*, 1965), but later as the late Eocene (Glezer, 1979).

In the section of the Tishki Formation near the village of Kantemirovka, Strel'nikova (1992) identified successive assemblages containing *Coscinodiscus* aff. *tennerimus*, *Hemiaulus polymorphus* v. *charkovianus*, *Coscinodiscus succinctus*, and, at the base of the Kas'yanovka Formation, an assemblage with *Rutilaria lymoniformis*. These assemblages were accepted as zonal, traceable in other sections, including oceanic, where (site 605 in the Atlantic Ocean) by nannoplankton they correspond to the zones NP15 - 16 of the mid-

Table 1. Conceptual changes of volumes of the Kiev and Khar'kov formations

Series	Subseries	Stage	Zone of planktonic foraminifers, Crimean-Caucasian scale	Horizons	Zosimovich *et al.*, 1966	Kraeva, 1974; Kraeva and Lyul'eva, 1977	Grigyalis *et al.*, 1988; Makarenko *et al.*, 1987	Scale in Report by GPA "Tsentrgeologiya" 1982, compiled by Iosifova and Shuleshkina	In present paper
Oligocene	lower	Rupelian		Khadum	Khar'kov Formation	Khar'kov Formation	Khar'kov Group: Mezhigorka Formation	Khar'kov Group: Kantemirovka Formation	
Eocene	upper	Priabonian	*Globigerapsis tropicalis*	Belaya Glina	Kiev Formation: Upper Subformation	Kiev Formation: Upper Subformation	Khar'kov Group: Obukhovo Formation	Khar'kov Group: Derizovka Formation	
	middle	Bartonian	*Globigerina turcmenica*	Kuma	Kiev Formation: Lower Subformation	Kiev Formation: Lower Subformation	Kiev Formation	Kiev Formation	Derizovka Formation
			H. alabamensis						
		Lutetian	*Acarinina rotundi-marginata*	Keresta		Buchak Formatoin		Buchak Formatoin	Kiev Formation
				Kuberli					
			A. bullbrooki		Buchak Formatoin		Buchak Formatoin		Buchak Formatoin



Localities of boreholes and their sections; intervals with carbonate and siliceous plankton assemblages; correlation with zonal bio-atratigraphic time scales; indices of transgressive-regressive cycles.

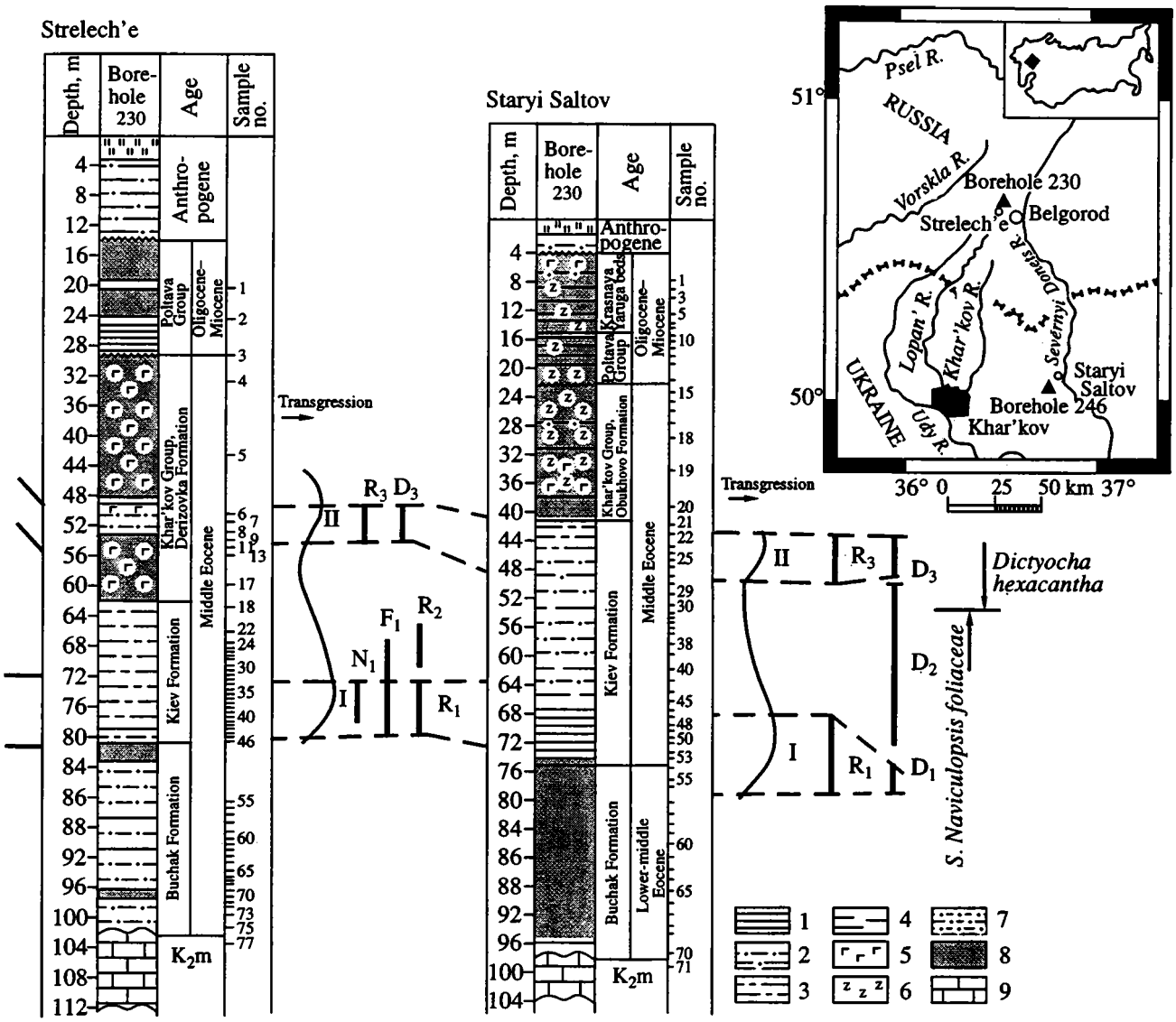
Transgression indices: I – Kiev transgression-regression cycle; II – Khar'kov trangression-regression cycle; Nl – nannoplanktonic assemblage of *Discoaster bifax* Zone; Fl – planktonic foraminifers assemblage of *Pseudoclavulina subbotinae-Hopkinsina bykovae ucrainica*. Radiolarian zones: R_1 – *Heliodiscus quadratus*; R_2 – *Cyrtoformis alfa*; R_3 – *Ethmosphaera polygiphonia*. Diatoms zones: D_1 – *Coscinodiscus tennerimus*; D_2 – *Hemiautus polymorphys* var. *charkovianus*; D_3 – *Coscinodiscus succinotus*; S – *silicoflagellates*. 1 – clays; 2 – sandy clay; 3 – silty clays; 4 – clay admixture; 5 – glauconite; 6 – mica; 7 – silts; 8 – sands; 9 – limestones.

dle Eocene (see the figure). The utility of Strel'nikova's scale is discussed below.

OBSERVATION MATERIALS

Our study concentrated on the lithology and microfossils from the sections of the boreholes 230 (Strelech'e) and 246 (Staryi Saltov) drilled by the Belgorod Geological Production Association (GPA) in 1990 - 1991 on the southeastern slope of the Voronezh syneclise near Belgorod (see the figure). Nazarkov described the sections and subdivided the rock sequence using the lithostratigraphic time scale elaborated by the GPA "Tsentrgeologiya" for the Paleogene deposits of the Voronezh anteclise on the basis of a modified Paleogene lithostratigraphic scale of the Ukraine (Makarenko *et al.*, 1987). Iosifova and Shuleshkina, in their unpublished report for 1982, subdivided the Eocene sequence into the Buchak, Kiev, and Derizovka (Obukhovo) formations. Three members were distinguished in the transgressive Kiev Formation: the lower sandy unit, the middle one of carbonate, siliceous or clayey sediments, and the upper member of clayey or siliceous deposits. The Khar'kov Group range in the Voronezh anteclise was suggested to include the Derizovka (Obukhovo), Pasekovo, and Kantemirovka formations. Only the lowermost of these formations, composed of glauconite sands and clayey silts, is dated as the Eocene.

In the sections of both boreholes (see the figure), the Paleogene rocks overlie the light-gray chalky limestones of the Maestrichtian, the erosion surface of which is covered by dark green and greyish-green mica-glauconite silts of the Buchak Formation. There are apparent scour marks at the base of the Kiev Formation. Its section begins with a sequence of light-colored sands (2 - 4 m) that grade into grey and greyish-green clays and clayey silts with small lenses and interbeds of quartz sand, which are 17 m thick in borehole 230 and 34.6 m thick in borehole 246. The Derizovka (Obukhovo) Formation is composed of yellow, grey, and greenish sands, which are micaceous-glauconitic or clayey-silty in several intervals and have 13.7-m thickness in borehole 230 and 17.6-m thickness in borehole 246. In both boreholes, the Obukhovo Forma-



The figure. (Contd.)

tion is overlain by the Neogene sands of the Poltava Group.

MICROPALEONTOLOGICAL ANALYSIS

Borehole 230 (Strelech'e)

Foraminifers. Two foraminiferal assemblages were observed: the lower one in the Buchak Formation (interval 87.5 - 83.8 m) with poorly preserved *Globigerina* and *Acarinina* forms, and the upper one in the Kiev Formation (interval 78 - 78.8 m) containing an abundant and diverse assemblage of almost 70 species of benthic foraminifers. The taxonomic composition and abundance of foraminifers from the lower part of the Kiev Formation are shown in Table 2. Representatives of the three genera *Haplophragmoides*, *Textularia*, and *Clavulina* with sandy shells and the primitive sandy foraminifers of the genus *Rhabdammina* were distinguished among the benthic foraminifers; those with calcareous (secretory) shells are more abundant and diverse than the agglutinated forms. Their most abundant genera are *Nodosaria* and *Guttulina* (nodosariids and polymorphinids), along with representatives of *Asterigera* and *Cibicidoides* (discorbidas and anomalinacids), *Hopkinsina, Uvigerina*, and *Bolivina* (buliminids).

Two types of paleocoenoses are distinguished. One of them contains nodosariids, rotaliids, and buliminids with a prevalence of diverse secretory foraminifers; the other includes rabdomminids, haplofragmoids, and clavulinids. The latter assemblage is represented only by shells with sandy walls. The coenoses of the first type contain highly diverse bionomic groups and occur in the interval of 78 - 72.5 m. The sandy foraminifers, dominant in the paleocoenosis of the interval 72.5 - 71.6 m, indicate deteriorated habitat environments for these organisms.

The stratigraphic and geographic distribution of some typical and index species of benthic foraminifers of the Kiev Formation, widely occurring in the European paleobiogeographic province, is shown in Table 3. Many of these species are present in the Lutetian deposits of Belgium (Kaasschieter, 1961) in the Bruxellian, Ledean, Wemmelian, and Assean deposits. Certain species appeared in the Ypresian, but most of them are found only in the upper half of the Lutetian and in the lower Bartonian of Belgium. The species of the studied assemblage are frequent in the Kuberli and Keresta horizons of the Crimea–Caucasus area and in the eastern part of Central Asia (Shutskaya, 1970; Nikitina, 1963, 1972; Kraeva and Zernetskii, 1969; Bugrova, 1988; Bugrova *et al.*, 1988). The assemblage corresponds to the *Pseudoclavulina subbotinae–Hopkinsina bykovae ucrainica* Zone (shown in the figure by F1), though its boundaries in this section are indeterminable. According to the benthic foraminifers, the studied deposits correspond to the *Hopkinsina bykovae ucrainica* Zone, observed in the Black Sea depression of the southern Ukraine (Kraeva and Zernetskii, 1969), to the beds with *Pseudoclavulina subbotinae* on the Scythian platform (Nikitina, 1972), and to the *Pseudoclavulina subbotinae–Bulimina granjuscula* Zone in the eastern part of the Caspian province (Ben'yamovskii *et al.*, 1992b). The first zone corresponds to the Kuberli Horizon, the second and the third to the Keresta Horizon.

Single finds of planktonic foraminifers *Pseudohastigerina micra* and *Globigerina turcmenica* are important for the determination of the age of the zone, because the former is common to the Lutetian and Bartonian and the latter is an index species of the *Globigerina turcmenica* Zone of the lower half of the Barhonian.

Nannoplankton. A diverse nannoplanktonic assemblage of more than 20 species, studied by Muzylev in the interval 79.6 - 73.0 m (shown in the figure by N1), includes *Coccolithus eopelagicus* (Braml. et Riedel), *C. formosus* (Kapt.), *Reticulofenestra umbilica* (Levin), *Rhabdosphaera gladius* (Loeker), *Discoaster strictus* Str., *Chiasmolithus solithus* (Braml. et Sull), and others. The assemblage is typical of the deposits with nannoplankton in the *Discoaster bifax* Subzone of the *Reticulofenestra umbilica* Zone in the Keresta Horizon and at the base of the Kuma Horizon in the southern CIS.

Radiolarians. The interval 79.5 - 78 m contains an abundant and taxonomically diverse assemblage of radiolarians with substituted siliceous skeletons. It has numerous representatives of the family Liosphaeridae, mainly *Cenosphaera* sp., *Carposphaera* sp., and *Thecosphaera* sp. There are also abundant *Prunopyle* sp., *Tristylosphaera* sp., *Thecotyle* sp., and taxa of the family Discoidea (*Spongodiscus* sp. and *Porodiscus* sp.). The generic composition, therefore, merely indicates the Paleocene–Eocene age of the assemblage.

The interval 76.5 - 75.4 m produced a profuse and diverse assemblage (shown in the figure by R_1), called the *Axoprunum visendum* assemblage after one of its typical species. The most ample and best preserved assemblage was recovered from a depth of 76.5 m where the representatives of the families Sphaeroidea, Prunoidea, and Discoidea, occurring together with nassellarians, dominate. The most abundant are *Spongotrochus paciferus* (Kozl.), *Thecosphaera* aff. *obliterata* (Zagorodnyk), *Heliodiscus heliasteriscus* (Comb. et Clark), and *Cromyodruppa tebegensis* (Lipman). Quite common are *Axoprunum* (*Ellipsoxiplus*) *visendum* Kozl., *Spongodiscus* aff. *alvaatus* (Riedel et Sanfilippo), and *Sethocytris babylonis* Riedel et Sanfilippo.

The species *Spongotrochus paciferus* and *Heliodiscus heliasteriscus* are almost universally present in the middle Eocene deposits, especially in the second half of the interval. *Axoprunum visendum* and *Cromyodruppa tenensis* are the typical middle Eocene species in provinces along the Volga River and in many regions of Middle Asia. In the lower latitudes, *Sethocytris babylonis* is confined to the middle part of the middle Eocene (the *Thyrsocytris triacantha–Podocyrtis mitra* zones of the tropical radiolarian scale), i.e., to the upper Lutetian Stage. In the Ypresian, *Axoprunum vigendum* appeared in the *Petalospyris fiscella* Zone (Kozlova, 1990) and existed till the end of the *Heliodiscus hexasteriscus* Zone, which correlates, apparently, with the larger part of the second half of the Lutetian. Consequently, the age of the assemblage, in all probability, corresponds to the latest time of the Lutetian.

The interval 75.4 - 67.5 m contains a diverse radiolarian assemblage, which not as well preserved as the previous one. This is an assemblage with *Heliodiscus quadratus* (shown in the figure by R_2). The radiolarian skeletons are sometimes substituted and associate with numerous sponge spicules. The representatives of the same families and genera as in the interval 76.5 - 75.5 m dominate here and include the following detected species: *Spongotrochus radiatus* (Lipman), *Cromyodruppa ovatus* (Grigorieva), *Pseudostaurosphaera* (?) *perelegans* (Krasheninnikov), *Axoprunum* (*Ellipsoxiplus*) *chabakovi* (Lipman), *Spongasteriscus gorskii* (Lipman), *Rhodosphaera* aff. *lipmanae* Zagorodnyk, and *Heliodigcus* cf. *quadratus*. The species of the genus *Lychnocanium* are rather abundant, especially the forms *Lychnocanium teopodicum* (Lipman) and *L.* cf. *bellum* (Riedel et Sanfilippo). In the lower latitudes, *Spongotrochus radiatus* occurs through the four upper radiolarian zones (*Podocyrtis ampla–Podocyrtis goetheana*) of the middle Eocene of the tropical scale. Its stratigraphic distribution in the middle latitudes seems to be of a wider range. The typical middle Eocene species of the Kuma Horizon in southern Russia, the Aral area, and the Turanian plate are *Rhodosphaera lipmanae*, *Stylosphaera irinia*, *Ellipsoxiplus chabakovi*, *Spongoasteriscus gorskii*, and *Cromyodruppa ovata*. The Kuberli and, partly, the Kuma horizons contain *Pseudostaurosphaera* (?) *perelegans* and *Lychnocanium separatum* (Moksyakova, 1972; Krasheninnikov, 1960).

The *Lychnocanium separatum* is an index species for the zone of the same name in Kozlova's boreal scale (1990), now referred to as the Lutetian Stage. The assemblage with *Heliodiscus quadratus*, however, can-

Table 2. Foraminifers from the base of the Kiev Formation

Foraminiferal community of the lower part of the Kiev Group (Borehole 230, interval 78 - 68.8 m)			
Benthic		*Guttulina* sp.	x
Order Astrorizida		*Globulina gibba* d'Orbigny	x
Rhabdammina eocenica Cushman et Hanna	x	*G.* ex. gr. *gibba* d'Orbigny	x - 0
Psammosphaèra fusca Schultze	x	*Oolina aplicata* d'Orbigny	x
Order Ammodiscida		Order Rotaliida	
Cornuspira ivolensis (Reuss) (= *Ammodiscus incertus* (d'Orbigny)		Superfamily Discorbacea	
Haplophragmoides rotundidorsatum Hantken	x	*Discorbis ergenensis Morozova*	x
H. glomeratum (Brady)	x	*Gyroidina octocamerata* Cushman et Hanna	x - 0
H. eggeri Cushman	x - 0	*Eponides umbonatus* (Reuss) (=*E. praeumbonatus* Mjatliuk)	x
Family Textulariidae		*Alabamina* ex. gr. *almaensis* (Samoilova)	x
Spiroplectammina pischvanovae A.V. et K.B. Furssenko	x - 0	*Asterigerina lucida* Minakova	0
S. guembelli Hagn. (= *Textularia labiata* Reuss)	x - 0	Superfamily Anomalinacea	
Textularia hueri d'Orbigny	x	*Cibicidoides* sp.	x - 0
T. flexibilis Kaptarenko	x - 0	*C. biumbonatus* A.V. et K.B. *Furssenko*	0 - 0
T. calix Grzybowski	x	*C. karpaticus* Mjatliuk	x - 0
Frankenia crassa Kaptarenko	x	*Brotzenella acuta acuta* (Plummer)	x
Order Ataxophragmiidae		*Heterolepa bionus* Schutzkaja	x
Gaudryina asiphonia (Andrea)	x	*H. pymeus* (Hantken) = *H. sulzensis* (Herrman)	x
Clavulinoides szaboi Hantken	x - 0	*Planulina burlingtonensis* (Jenning) (= *Cibicides cribrosus* A. V. et K. B. Furssenko)	x - 0
Clavulina parisiensis d'Orbigny	x - 0	*Anomalina nonioninoides* A. V. et K. B. Furssenko	0 - 0
C. cylindrica Hantken	x - 0	*A. affinis* Hantken	x - 0
Pseudoclavulina subbotinae I. Nik = *P. terkensis* (Chalilov) = *P.* cf. *cocaensis* (Cushman)	x - 0	Family Nonionidae	
Order Nodosariida		*Nonion scaphum* Fichtel et Moll = *N. ovalis* (Chalilov)	x - 0
Family Nodosariidae		Order Buliminida	
Lagena laevis Montagu	x	*Bifarina millepunctata* (Tutkowski)	0
L. striata d'Orbigny	x	*Turrilina alsatica Andrea*	x - 0
L. lineata-punctata Heron-Allen et Earland	x	*Bulimina ovata* d'Orbigny	x
Nodosaria longiscula (d'Orbigny)	x - 0	*Uvigerina proboscidea Schwager* (= *U. farinosa* Hantken = *U. batjesi* Kaasciter (partly) = *U. garzaensis* Cushman et Siegfus)	0
N. guttifera (d'Orbigny)	x	*U. huspida* Schwager (= *U. garzaensis hudo-robusta* Mallory)	0
N. ewaldi Reuss	x - 0	*U. costellata* Morozova (= *U. churchi demicostata* Mallory)	0 - 0
Robulus sp.	x	*U. cocaensis* Cushman	x - 0
R. inornata (d'Orbigny)	x	*Hopkinsina bykovae ucrainica* Krajeva (= *Uvigerina spinocostata* Cushman et Jarvis)	0 - 0
R. calcar (Linne)	x	*Trifarina bradyi* Cushman (= *T. wilcoxensis* Cushman et Ponton)	x - 0
Dentalina inornata d'Orbigny	x	*Bolivina reticulatiformis* Chalilov (= *B. cookei* Cushman)	x - 0
Marginulina pseudosetosa Morozova	x	*B. quasiplicata* Subbotina (= *B. crenatula* Cusman)	x
M. deformis Reuss	x	*B. microplancetiformis* Subbotina = ? *B. pulchra* (Terquem)	x - 0
M. deformis spinulosa Sherborn et Chapman	x	*Furssenkoina* (Virgulina) *dibbolensis* Cushman et Applin	x - 0
Vaginulinopsis mexicana (Cushman)	x	Family Chilostomellidae	
V. decorata (Reuss)	x - 0	*Pullenia sphaeroides* (d'Orb.)	x
Family Polymorphinidae		*P. quiqueloba* (Reuss)	x
Guttulina pulchella d'Orbigny	x	Planktonic	
G. problema d'Orbigny	x - 0	*Globigerina turcmenica* Chalilov	x
G. irregularis d'Orbigny	x	*Pseudohastigerina micra* Cole	x

Note: x – single and rare finds up to 10 specimens, 0 – normally more than 10 specimens.

Table 3. Stratigraphic and geographic distribution of the typical species of foraminiferal assemblages in the lower Kiev deposits (Borehole 230; interval 80 - 68.8 m)

Taxa \ Regions	Southern European part of Russia and adjacent areas (Dnieper–Donets depression; Belarus'; the Volga provinces; the Caspian province; Crimea; Ciscaucasus)				Belgium				
Lithostratigraphics units	formations								
	Simferopol'	Kuberli	Keresta	Kuma	Ypresian	Bruxellian	Lede Formation	Wemmel Formation	Asse Formation
		Kiev							
Stages	Ypresian	Lutetian		Bartonian	Ypresian	Lutetian			Bartonian
Spiroplectammina pischvanovae									
Clavulina cylindrica									
Pseudoclavulina subbotinae									
Lagena striata									
Nodosaria ewaldi									
Dentalina inornata									
Vaginulinopsis decorata									
Guttulina pulchella									
Asterigerina lucida									
Anomalina affinis									
Cibicides cribrosus (= *Planulina burlingtonensis*)									
Cibicidoides biumbonatus									
Heterolepa bionus (= *Cibicides sulzensis*)									
Nonion scaphum									
Uvigerina proboscidea (= *U. farinosa* = *U. batjesi*)									
U. huspida (= *U. garzaensis*)									
U. costellata									
Turrilina alsatica									
Trifarina bradyi (= *T. wilcoxensis*)									
Hopkinsina bykoval ucrainica									
Bolivina reticulatiformis (= *B. cookei*)									
Globigerina turcmenica									
Pseudohastigerina micra									

not be attributed reliably to the *Lychnocanium separatum* Zone, because this index species appears in the overlying deposits as well. The composition of the association indicates, more likely, the early Bartonian age of the host deposits.

The interval 66.8 - 62.0 m contains a large amount of sponge spicules and poorly preserved radiolarians of the Liosphaeridae family.

The interval 56.5 - 49.0 m shows taxonomically diverse and almost intact radiolarians of the *Heterosestrum shabalkini* assemblage (shown in the figure by R_3): *Spongolissa callosa* Kozl., *Stylosphaera coronata coronata* Hek., *St.* aff. *rosula* Ehr., *Druppatractus santaennae* Camp. et Clark., *Haliomma rosula* Kozl., *Plectodiscus circularis* Ehr., *Dithomelissa haeckeli* Ehr., *Ommatogramma bilobatus, Heterosestrum* aff. *formosum* Kozl., and *H. shabalkini*.

The species, mainly boreal, of the assemblage were known earlier in the Norwegian basin, the Dnieper–Donets depression, the Volga provinces, and western Siberia. The species *Heterosestrum formosum* occurs in the Bodrak Stage of the Voronezh anteclise and the Dnieper–Donets depression (Tochilina, 1975; Zagorodnyuk, 1970) and was detected in the middle Eocene

assemblage of the Norwegian Sea together with *Phacodiscus testatus grandis* (Kozlova and Petrushevskaya, 1979). The species *Haliomma rosula* is typical of the presumably upper Eocene deposits of the Norwegian Sea, and *Spongomelissa callosa* was observed by Kozlova in the so-called RE-3 zone of the Volga provinces, now referred to the second half of the middle Eocene.

The age interval of the radiolarian species from the Norwegian basin is so far poorly known. The age determination of their host deposits is hindered by the absence of data on other planktonic groups and by the endemic character of radiolarians. The species *Stylosphaera coronata coronata, Druppatractus santaennae,* and *Plectodiscus circularis* evidently belong to the cosmopolitan species of a wide stratigraphic range (the middle-upper Eocene). The presence of *Heterosestrum shabalkini* and associated forms is recognized in the Isfara Beds of the Ferghana province and in the Tishki Horizon in the Voronezh anteclise (Tochilina, 1966; Kozlova and Petrushevskaya, 1979). The assemblage with *Heterosestrum shabalkini*, identified by the latter two authors in the Norwegian Sea, has no definite age correlation.

Diatoms and silicoflagellates. Abundant and excellently preserved diatoms and silicoflagellates are observed in the interval 56.9 - 49 m of the Derizovka Formation. The taxonomic composition of the assemblages is practically identical in all of the studied samples (D_3 in the figure shows the interval of the assemblage). The most numerous are *Paralia sulcata* var. *crenulata* Grun., *P. ornata* Grun., *P. clavigera* (Grun.) Gles., var. *concavus* Gles., and *Cosmiodiscus* sp. 1 similar to *Coscinodiscus subtilis* (sensu Stel'nikova, 1979, figures 55 and 56). Rather abundant and taxonomically diverse are *Hemiaulus* and *Pyxidicula* genera: *H. ambiguus* Jousé, *H. subacutus* Grun., *H. polymorphus* var. *frigida* Grun., *H. polycystinorum* Ehr., *Pyxidicula turris* (Grev. et Arnott) Streln. et Nikolaev, *P. charkoviana* (Jousé) Streln. et Nikolaev, *P. crenata* (Sheshuk.) Streln. et Nikolaev, and *P. corona* (Ehr.) Streln. et Nikolaev. The typical forms of the interval are: *Pseudopodosira westii* (W.S.) Sheshuk. et Gles., *P. pileiformis* Jousé, *Peponia barbadense* Grev., *Jousea elliptica* (Jousé) Gles., *Biddulphia tuomei* (Bail.) Roper, and *Rutilaria aveolata* Sheshuk. Other species are more scarce: *Asterolapra insignis* Grev., *Craspedodiscus* Ehr., *Coscinodiscus bulliens* A.S., *C. succinctus* Gles., *C. duplex* Gles., *Rhizosolenia* aff. *gravida* Gombos, *Pyxilla* aff. *prolongata* Brun., *Rhizosolenia clavigera* (Grun.) Homman, *Brightwellia coronata* (Bright.) Ralfs., and *B. spiralis* Gles.

The lithology of the host rock entirely corresponds to the ecological pattern of the assemblage containing dominant neritic genera and species, especially to the presence of *Paralia sulcata* and its varieties characteristic of sandy shoals. The presence of silicoflagellates and a few species of oceanic diatoms implies sedimentation in an open marine basin.

Setting aside the discussion of the taxonomic composition and morphological aspects of the defined diatom assemblages, we wish to note that the study with the scanning electron microscope revealed a significant deviation in the structure of the valve margins of *Coscinodiscus succinctus* and a similar species *C. duplex*, described by Glezer, from the typical features of the genus *Coscinodiscus*. Accordingly, all species of the genus *Cosmiodiscus* require revision and a more accurate definition, but unfortunately, the evolutionary stages of the Eocene diatom flora at the generic level are not known, and remain a subject for future research.

The stratigraphic position of the assemblage and, consequently, the age of the Derizovka Formation are determined from the data on the vertical range of individual species of diatoms and silicoflagellates, which correlate to standard scales based on the calcareous plankton. According to Perch-Nielsen (1985), the widespread presence of the silicoflagellate *Dictyocha hexacantha* outside the tropical regions of the World Ocean suggests that the studied level is between the end of the middle and the beginning of the late Eocene. The data on diatoms slightly narrows this stratigraphic interval. Rare species of *Coscinodiscus* (?) *succinctus* commonly found in the assemblage are the index species of the relevant zone corresponding to the end of the middle Eocene in Strel'nikova's scale (1992); this zone correlates to the upper *Discoaster saipanensis* Subzone of the *Reticulofenestra umbilica* Zone (= NP16).

The age of the Derizovka Formation, inferred from the siliceous plankton evidence, corresponds to the age of the underlying deposits of the Kiev Formation dated by Muzylev according to nannoplankton. Consequently, we referred the interval of 79.5 - 73.5 m in borehole 230 to the lower *Discoaster bifax* Subzone of the *Reticulofeneatra umbilica* Zone. Thus, the correlation of the Derizovka Formation with the upper subzone of *Reticulofenestra umbilica* Zone seems reliable.

In borehole 230, the rocks above the analyzed interval contain only single and redeposited specimens of radiolarians, foraminifers, and sponge spicules.

Borehole 246 (Staryi Saltov)

In contrast to borehole 230, this site completely lacks nannoplankton, and foraminifers in it are extremely rare, but the interval containing diatoms covers an obviously longer time period.

Diatoms and silicoflagellates. Diatoms are found in the interval of 77 - 42 m. The first specimens appear already at a depth of 77.0 m in the sand beds referred to the Kiev Formation. The assemblage contains *Coscinodiscus obscurus, C. obscurus* var. *concavus*, Gl., *Brightwellia johnsonii* Grev., and *Hemiaulus polymorphus* var. *charkovianus* Jousé. The species are similar to forms in the overlying assemblage.

The interval of 67.0 - 41.0 m contains abundant specimens of diatoms and silicoflagellates, the latter consisting of two assemblages. The first, deposited in the interval of 67 - 51 m, includes *Naviculopsis foliacea* Defl., *Corbisema triacantha* (Ehr.), and *Dictyocha subclinata Bukry.* The second, in the interval of 51 - 42 m,

contains *Dictyocha hexacantha* Schulz., *Distephanus crus* Ehr., and *Macrora barbadense* (Defl.) Bukry. Both assemblages have zonal species and can be easily correlated with the zones *Naviculopsis foliacea* and *Dictyocha hexacantha*, detected in the scales of Martini, Bukry, and Perch-Nielsen (Bolli *et al.*, 1985). The zonal species *Dictyocha hexacantha* appears at the base of the zone NP16, definitely correlating this interval with the base of the upper third part of the middle Eocene (see the figure).

Most of the more than 50 species of the diatom assemblage are widespread and long-lived neritic and oceanic forms. At first glance, the assemblage composition seems monotonous through the entire interval of the section, consisting of *Paralia sulcata* Ehr., *P. oamaruensis* (Gr. et St.), *Melosira ornata* (Grun.), *Pyxilla gracilis* Témp. et Forti, *P. oligocenica* var. *tenius* Jousé, *Hemiaulus polycystinorum* Ehr., *H. polymorphus* var. *charkovianus* Jousé, *Coscinodiscus bulliens* A.S., and *C. obscurus* Ehr. Nevertheless, the appearance of index species, sometimes as single specimens, still makes it possible to subdivide the assemblage. The interval of 77 - 66 m contains *Coscinodiscus* aff. *tennerimus* Jousé along with *Brightwellia johnsonii* Grev. and *Hemiaulus proteus* Ehr., which are lacking in the higher horizons. This assemblage can be referred to the *Coscinodiscus* aff. *tenerrimus* beds (D_1 interval in the figure) of Strel'nikova's scale (1992).

The interval of 62 - 52 m includes an assemblage corresponding to the *Hemiaulus polymorphus* var. *charkovianus* beds (D_2) of the same scale. The upper boundary of the assemblage is detected by a considerable reduction in the amount of zonal species valves, and by the disappearance of *Coscinodiscus obscurus* var. *concavus* and *Pyxilla gracilis*.

The upper assemblage (D_3; interval 52 - 42 m) corresponds to the *Coscinodiscus succinctus* beds of Strel'nikova's scale, and to the *Triceratium inconspicuum* var. *trilobata* Zone of Barron's scale (McLean and Barron, 1989). The characteristic feature of this assemblage is the appearance of *Coscinodiscus succinctus* and *Hemiaulus subacutus* and a greater abundance of *Triceratium inconspicuum* var. *trilobata* Fenner and *Rhizosolenia* aff. *gravida* valves.

The composition of the neritic species in the interval is more rich at the expense of more abundant forms of such genera as *Corona* (C. *retinervis* Sheshuk. et Gles.) and *Aulacodiscus* (*A. instans* Grev., *A. kelleri* var. *fascicularis* Sheshuk. et Gles.) supplemented with *Brightwellian spiralis* (Brign.) Gles., *Hyalodiscus scoticus* Ehr., and *Pseudopodosira hyalina* Jousé. The data on diatoms entirely coincide with the data on silicoflagellates, i.e., the identified assemblages correspond to the nannoplankton zones NP 15 - 16. Typical species of diatoms and silicoflagellates recovered from boreholes 230 and 246 are shown in Plates I - V.

Radiolarians. The *Heliodiscus quadratus* assemblage (R_1).

The interval of 75.2 - 69.3 m contains rare radiolarians mainly represented by the Discoidea (especially Spongodiscidae) species *Axoprunum chabakovi* (Lipman), *Spongotrochus radiatus* (Lipman), *Heliodiscus* aff. *quadratus* Camp. et Clark., *Heliodiscus heliasteriscus* (Clark et Campbell), *Porodiscus circularis* Clark et Campbell, and *Spongodiscus communis* Lipman. The last two species are of wide geographic distribution and probably occurred throughout the entire Eocene–Oligocene. The first two species are typical of the second half of the middle Eocene. In the lower latitudes, *Spongotrochus radiatus* is frequently observed in the radiolarian *Thyrsocyrtis triacantha–Podocyrtis mitra* zones (Khokhlova's data on the Bay of Biscay and Cyprus). *Heliodiscus quadratus* is the index species of the relevant zone in the lower Bartonian deposits of the European part of Russia (the foraminiferal *Hantkenina alabamensis* Zone). *Heliodiscus heliasteriscus* Camp. et Clark. is common for the middle Eocene and *Axoprunum chabakovi* for the upper part of the middle Eocene (the Kuma Horizon) in the Aral area and Turanian plate. Therefore, this assemblage is most likely early Bartonian in age.

The *Heterosestrum shabalkini* assemblage (R_3).

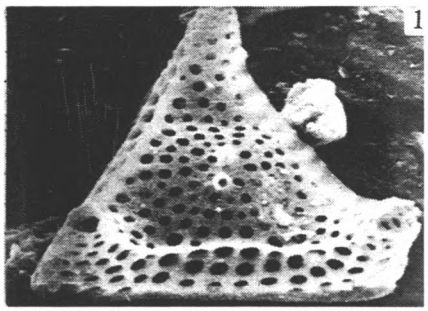
In the interval of 53 - 43 m, radiolarians are represented by *Stylosphaera coronata coronata* Hek., and *Pectodiscus circularis* Clark et Campbell, which are species with a wide stratigraphic and geographic distri-

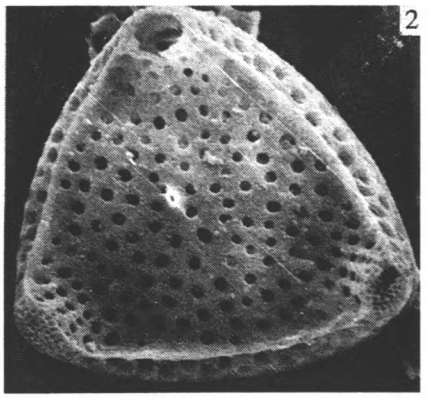
The Middle Eocene diatoms from the Khar'kov and Derizovka formations of the northeastern slope of the Dnieper–Donets depression (Boreholes 230 and 246; Belorussian Geological Prospecting Expedition).

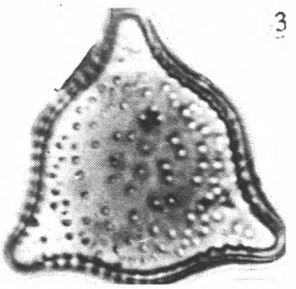
Plate I. 1, 2, 4, – *Triceratium exornatum* Grev., 1, 2 – borehole 230, sample 8, ×1500; 4 – borehole 230, sample 13, ×1500. 3 – *Triceratium sparsipunctata* Jousé, borehole 230, sample 9, ×1000. 5 – *Trinacria subcapitatum* (Grev.) Grun, borehole 230, sample 13, ×1500. 6 – *Trinacria excavata* Heib., borehole 230, sample 8, ×750. 7 – *Trinacria excavata* f. *tetragona* A.S., borehole 230, sample 7, ×1000. 8 – *Jousea elliptica* (Jousé) Gles., borehole 230, sample 13, ×3500. 9, 10 – *Rhizosolenia clavigera* (Grun.) Homman, 9 – borehole 246, sample 24, ×1000; 10 – borehole 246, sample 42, ×1000.

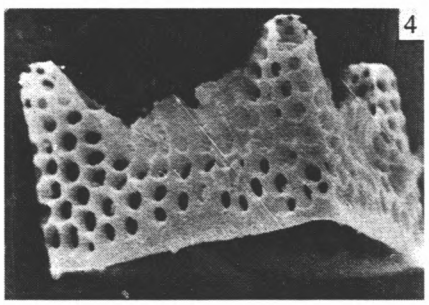
Plate II. 1, 2 – *Hemiaulus polymorphus* var. *charkovianus* Jousé, borehole 246, sample 24, ×1500. 3 – *Hemiaulus subacutus* Grun., borehole 230, sample 8, ×1500. 4, 8 – *Hemiaulus ambiguus* Jousé, borehole 230, sample 8, ×1500. 5 – *Hemiaulus polycystinorum* Ehr., borehole 230, sample 8, ×1500. 6 – *Hemiaulus* aff. *polymorphus* var. *frigida* Grun., borehole 246, sample 24, ×1500. 7 – *Biddulphia tuomeyi* (Bail) Roper var. *charkoviana* Jousé, borehole 246, sample 8, ×1500. 9, 10, 11 – *Dictyocha hexacantha* Schulz, 9, 10 – borehole 230, sample 13, ×1000; 11 – borehole 230, sample 8, ×1500; 12 – *Naviculopsis foliaceae* Deflandre, borehole 246, sample 24, ×750.

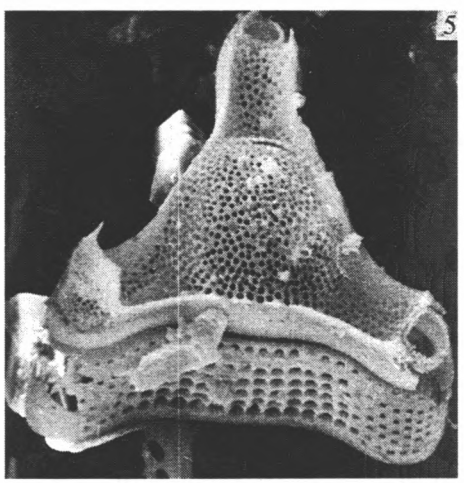
Plate III. 1, 5 – *Pyxidicula charkoviana* (Jousé) Streln. et Nikolaev, borehole 246, sample 24, ×1500. 2, 3, 4 – *Pyxidicula turris* (Grev. et Arnott) Streln. et Nikolaev, 2 – borehole 230, sample 8, ×750, 3, 4 – borehole 246, sample 24, ×1500. 6 – *Pyxidicula* aff. *corona* (Ehr.) Streln. et Nikolaev, borehole 230, sample 8, ×1000. 7, 8 – *Pyxidicula* sp.1, borehole 246, sample 30, ×1000. 9 – *Omphalotheca* aff. *jutlandica* Grun. em. Homann, borehole 230, sample 13, ×1500. 10 – *Pseudostictodiscus ovetschkinii* Gles., borehole 246, sample 24, ×1000. 11 – *Pyxidicula cruciata* Ehr., borehole 246, sample 24, ×1000.

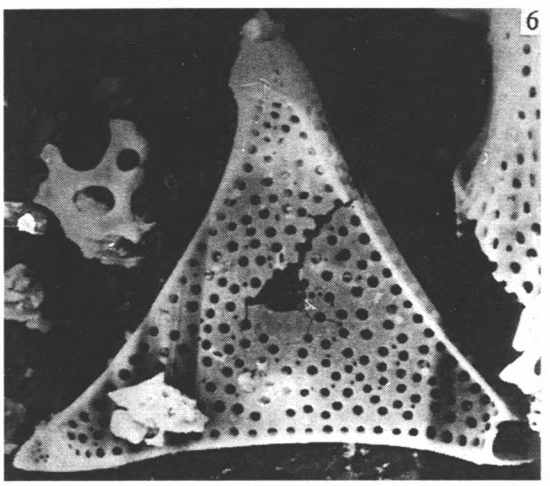


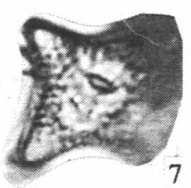


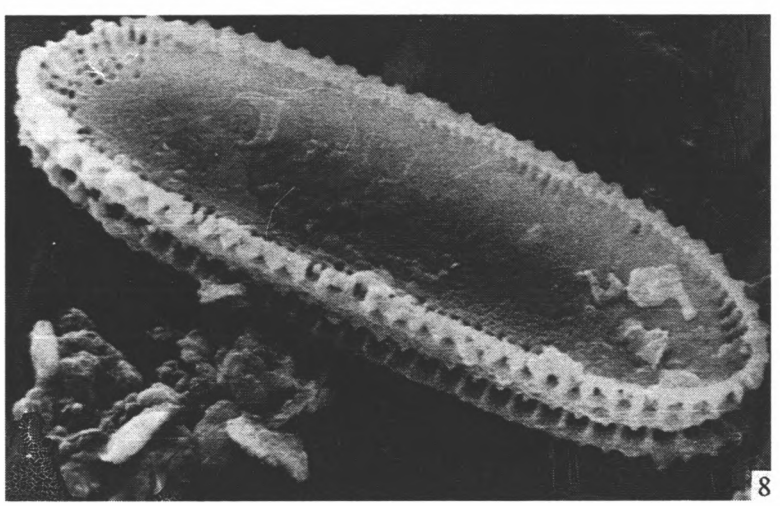












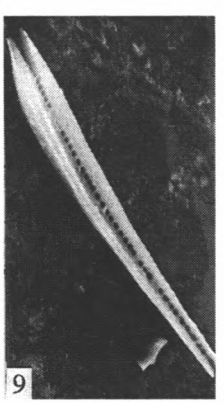
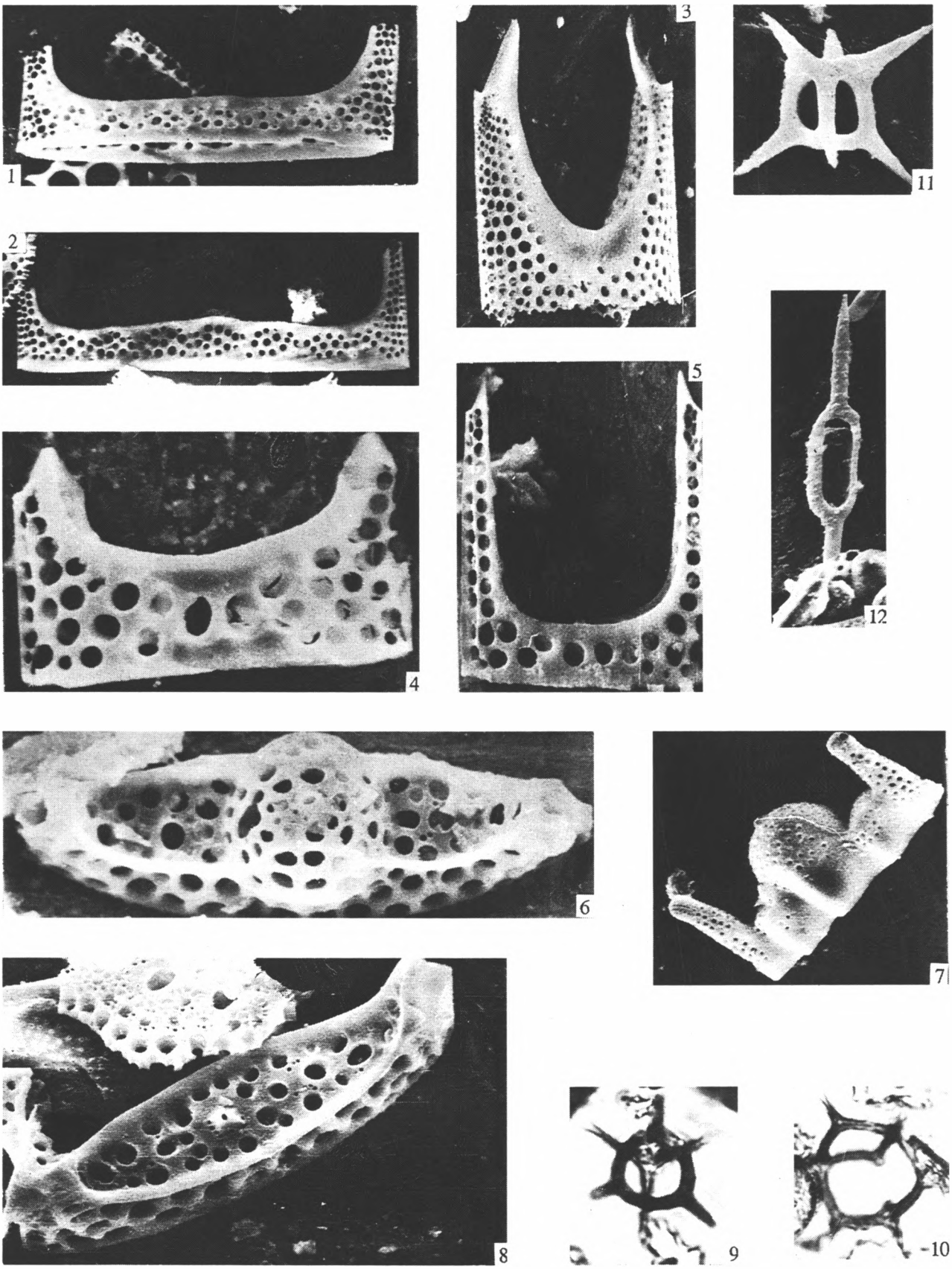


Plate I.



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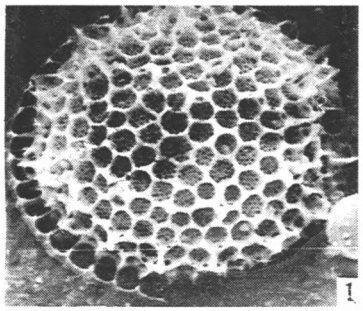
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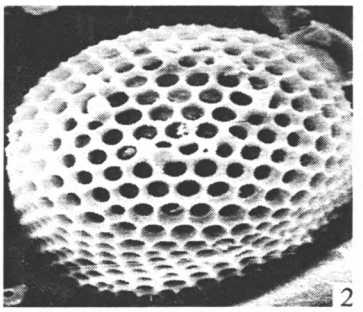
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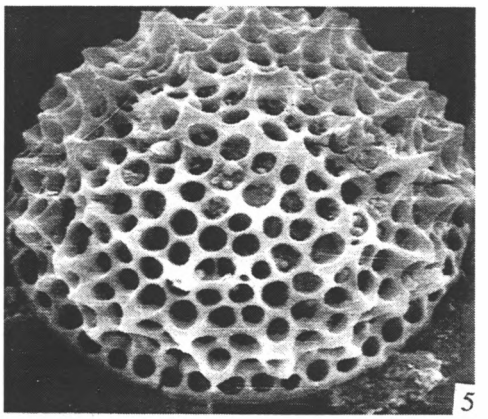
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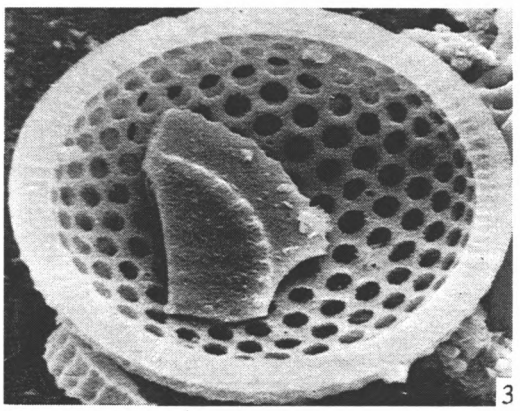
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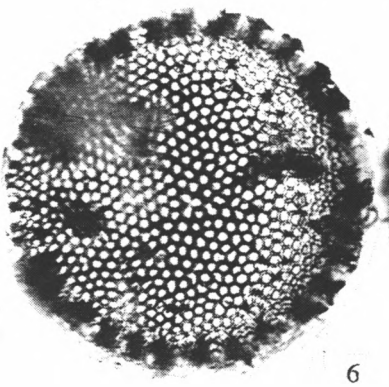
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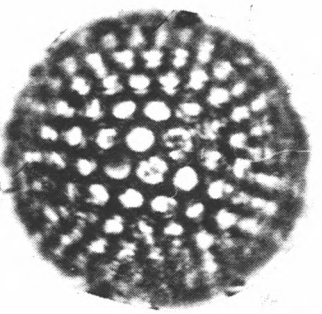
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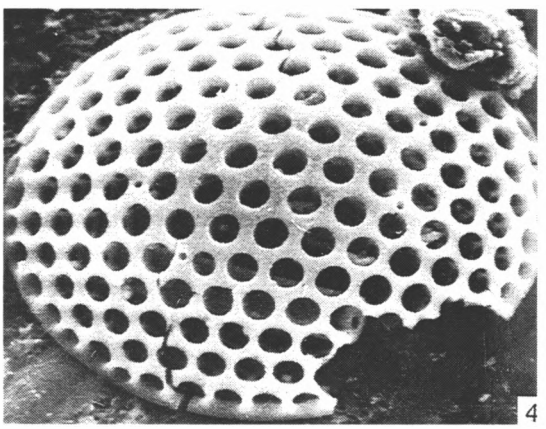
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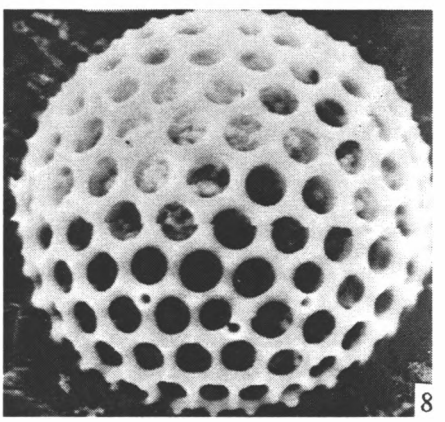
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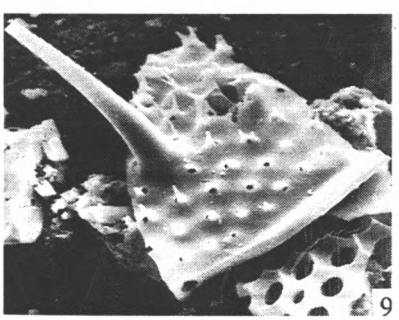
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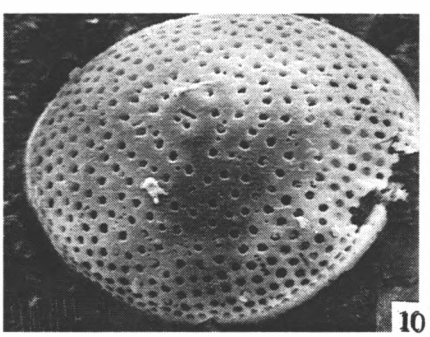
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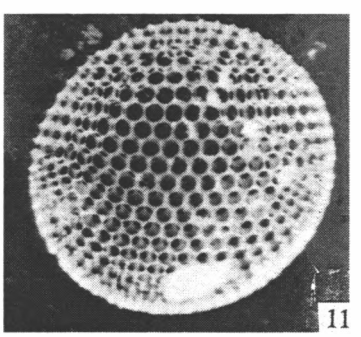
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Plate III.

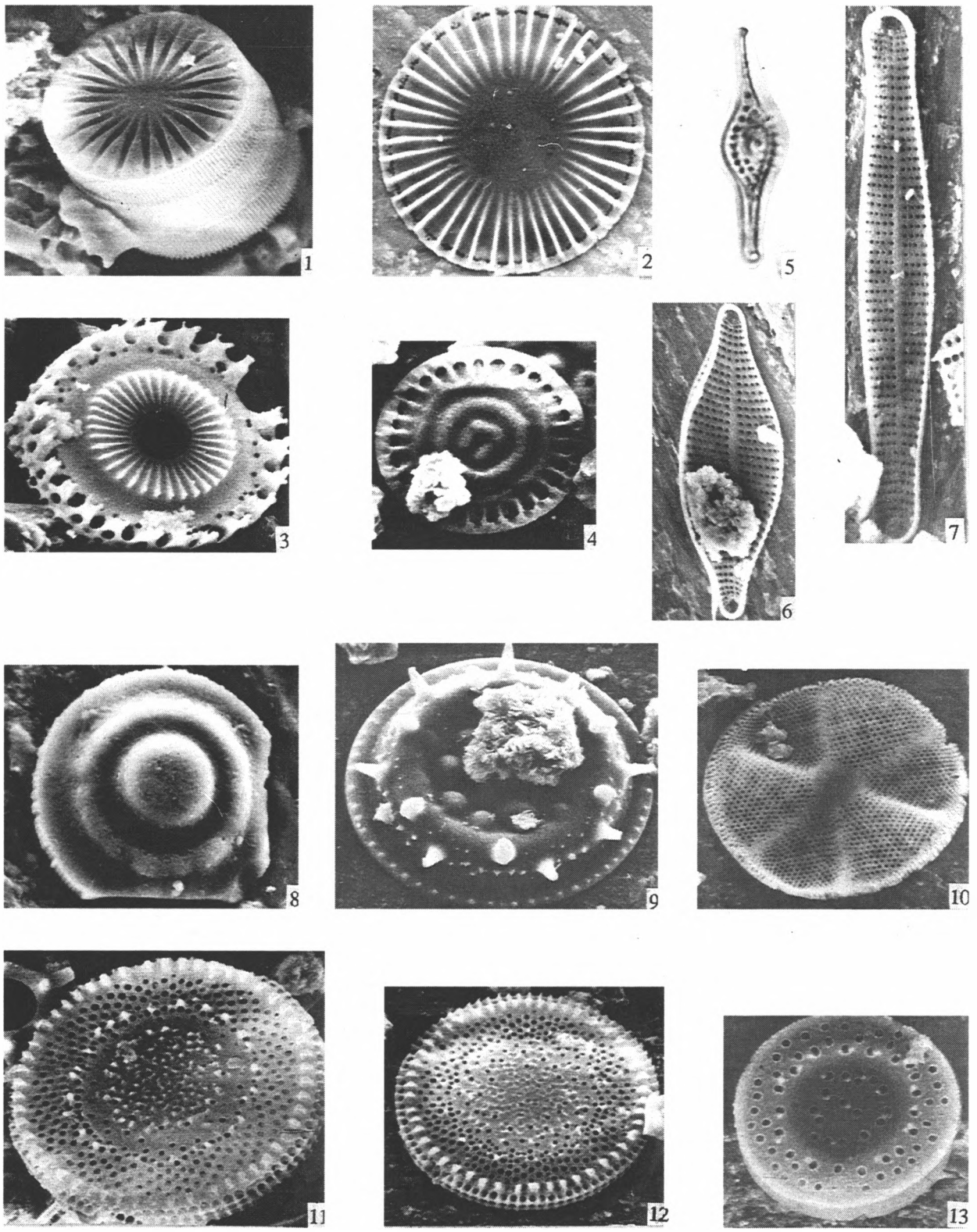
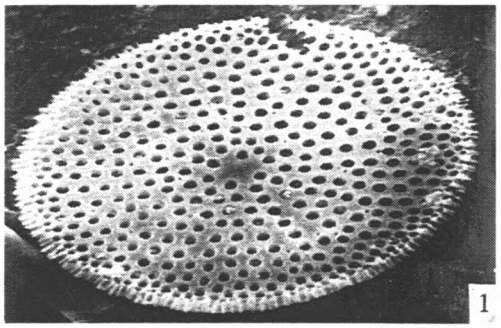
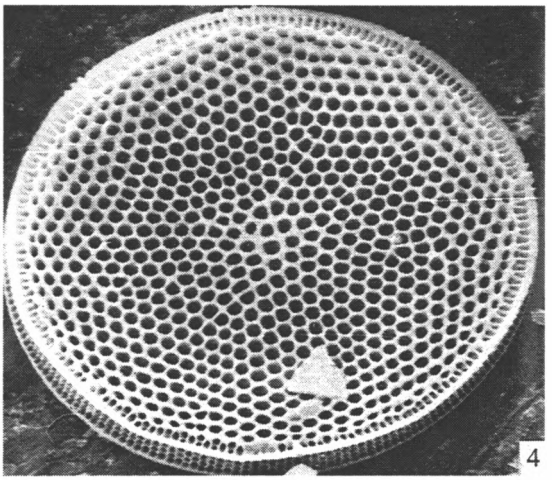
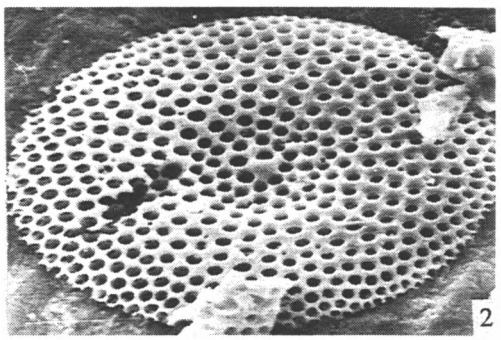
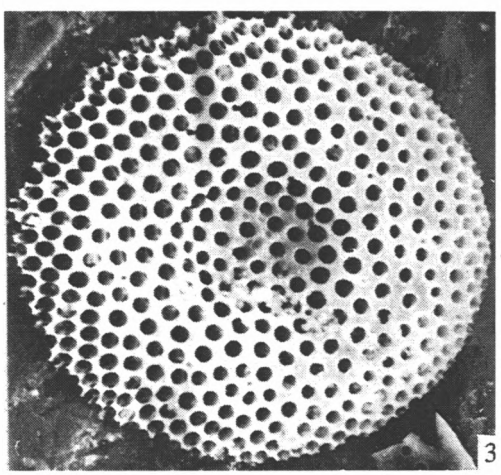


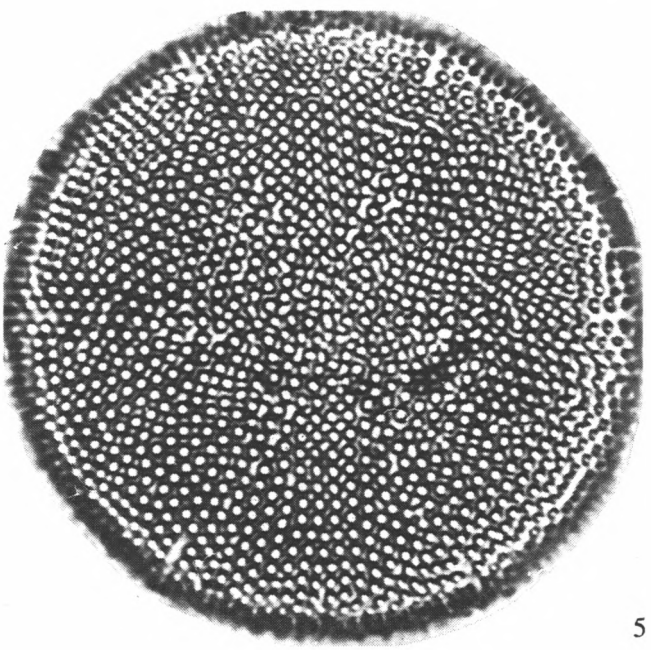
Plate IV. 1, 2 – *Paralia clavigera* (Grun.) Gles., 1 – borehole 230, sample 8, ×750; 2 – borehole 246, sample 24, ×1600. 3 – *Paralia sulcata* var. *crenulata* Grun., borehole 230, sample 8, ×1500. 4 – *Paralia ornata* Grun., borehole 230, sample 8, ×1400. 5 – *Rutilaria areolata* Sheshuk., borehole 230, sample 7, ×1500. 6 – *Rhaphoneis ampiceros* Ehr., borehole 230, sample 13, ×1500. 7 – *Rhaphoneis simbirskiana* Pant., borehole 230, sample 8, ×1500. 8 – *Pseudopodosira westii* (W.S.) Sheshuk. et Gles., borehole 230, sample 8, ×1500. 9 – *Pseudoporosira pileiformis* Jousé, borehole 230, sample 8, ×1500. 10 – *Actinoptychus senarius* Ehr., borehole 230, sample 13, ×1500. 11, 12, 13 – *Melosira architectularis* Brun., 11, 12 – borehole 230, sample 8, ×1500; 13 – borehole 230, sample 13, ×1500.

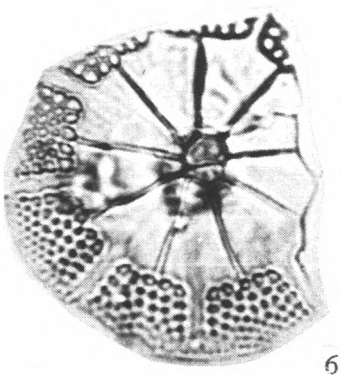


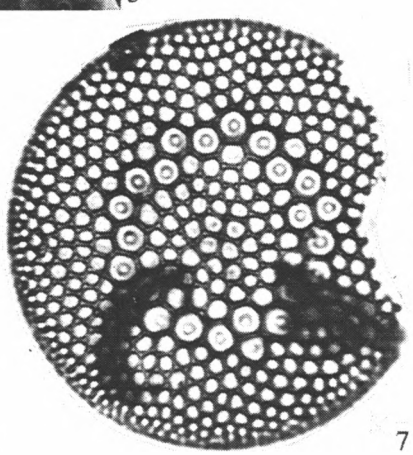












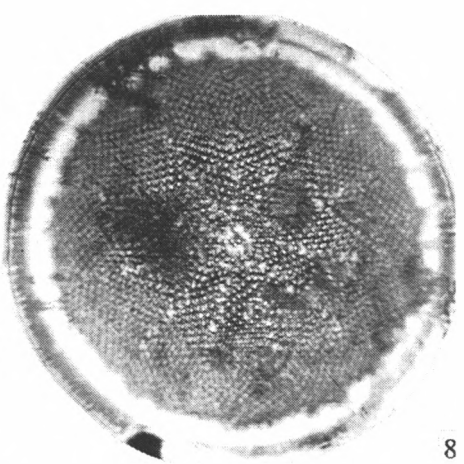


Plate V. 1 – *Coscinodiscus obscurus* A.S., borehole 246, sample 24, ×750. 2, 3 – *Coscinodiscus obscurus* var. *concavus* Gles., borehole 246, sample 24, ×750. 4 – *Coscinodiscus* (?) *duplex* Gles., borehole 230, sample 13, ×1500. 5 – *Cosmiodiscus* sp.1 (aff. *Coscinodiscus subtilis* Ehr. *sensu* Strel'nikova, 1979, Figs. 55 and 56), borehole 230, sample 13, ×1500. 6 – *Asterolapra decora* Grev., borehole 230, sample 13, ×1000. 7 – *Coscinodiscus bulliens* A.S., borehole 230, sample 13, ×1000. 8 – *Coscinodiscus* (?) *succinctus* Gles., borehole 230, sample 13, ×500.

bution. Some species of the second half of the middle Eocene, observed in the Volga provinces, western Siberia, and Turkmenistan, are also present: *Petalospyris tumidula* Kozl., *Hexacontium micrum* Lipman, *Dictyospyris* (?) *tetrastoma* (Ehr.), *Druppatractus santaennae* Camp. et Clark., and *Carposphaera* aff. *magnoporulosa* Lipm. Two specimens of *Heterosestrum* aff. *shabalkini* Lipman were also found. This species was observed in the southern Russian plain and in the Isfara Beds of the Middle Asia, which belong to the upper unit of the middle Eocene (Lipman, 1975). In the Norwegian–Greenland basin, *Heterosestrum shabalkini* is the typical species of the assemblage whose stratigraphic position, in all likelihood, is also of the same level as the middle Eocene (Kozlova and Petrushevskaya, 1979).

In spite of the relative taxonomic paucity of the assemblage, it is, in many ways, similar to the R_3 assemblage from borehole 230, and the presence of *Heterosestrum shabalkini* suggests its late Bartonian age.

RESULTS AND DISCUSSION

A sequence of carbonate silts in borehole 230 contains diverse foraminifers (79.6 - 69.8 m) and nannoplankton (79.6 - 73.0 m) of late Lutetian–early Bartonian age. According to benthic foraminifers, the sequence was identified as the *Pseudoclavulina subbotinal–Hopkinsina bykovae ucrainica* Zone, whose range corresponds to that of the *Hantkenina alabamensis* Zone and to the lower part of the *Globigerina turcmenica* Zone of the planktonic foraminifers. According to the nannoplankton, the sequence is equivalent to the *Discoaster bifax* Subzone of the *Reticulofenestra umbilica* Zone and, as a rock unit, corresponds to the lower half of the marl beds of the Kiev Formation, or to the Keresta and the base of the Kuma horizons in more southern regions.

Three radiolarian assemblages are identified in boreholes 230 and 246. The lower assemblage with *Axoprunum vigendum* and the middle one with *Heliodiscus* cf. *quadratus* characterize the Kiev Formation, and the upper *Heterosestrum shabalkini* assemblage, recovered from borehole 246, corresponds to the lower part of the Derizovka Formation. In borehole 246, however, the upper assemblage occurs in the sandy clays of the top level of the Kiev Formation, as was determined by field research. The deposits bearing two lower assemblages correlate to the upper part of the *Lychnocanium separatum* Zone and with the lower part of the *Heliodiscus quadratus* Zone, according to Kozlova's scale (1990). The deposits containing the upper assemblage correspond to the upper part of the *Cythoformis* alta Zone and the lower part of the *Ethmosphaera polysiphonia* Zone on the same scale. Therefore, the two lower beds with radiolarians can be dated as the late Lutetian–early Bartonian, and the upper unit as the middle Bartonian.

The most clearly defined are the stratigraphic boundaries determined according to the distribution of silicoflagellates. In borehole 246, we recognizeed boundaries of the *Dictyocha spinosa–Dictyocha hexacantha* zones, and in borehole 230, only the *Dictyocha hexacantha* Zone was identified.

Among the three diatom assemblages, the lower one with *Coscinodiscus* aff. *tenerrimus* was identified only in borehole 246 at the base of the Kiev Formation. Higher up above this assemblage, in the same borehole, we detected an assemblage with *Hemiaulus polymorphus* var. *charkovianus*, and a third group with *Coscinodiscus succinctus* was determined at the top of the Kiev Formation in borehole 246 and at the base of the Derizovka Formation in borehole 230.

The composition of diatoms, radiolarians, and silicoflagellates in the assemblages permits the correlation of the upper part of the Kiev Formation (borehole 246) and the lower part of the Derizovka Formation (borehole 230) with the base of the *Dictyocha hexacantha* silicoflagellate zone, with the *Coscinodiscus succictus* diatom zone, and with beds the containing *Heterosestrum chabalkini* according to radiolarians. All of these subdivisions are referred to zone NP 16, i.e., to the Bartonian. Thus, the Derizovka Formation, which is an analog of the Obukhovo Formation and the lower unit of the Khar'kov Group, is of middle Eocene age.

The figure represents the initial subdivision of layers, i.e., the upper part of the Kiev Formation in borehole 246 and the base of the Khar'kov Formation in borehole 230 are shown having similar faunal characteristics. This error, as we believe, is not accidental being a result of the assumed lithologic homogeneity of the Khar'kov Formation which lacks regular changes in rock facies, as well as reflecting the notion that there are distinct lithological differences between the Kiev and Khar'kov sedimentary sequences. These were reliable assumptions concerning the deeper central part of the Dnieper–Donets depression, but on its slopes, the facies and lithology of deposits change, i.e., the Kiev Formation becomes more terrigenous and siliceous, and the differences between the formations are less evident.

Two transgressive-regressive cycles in the rock sequences are recognized according to analyzed changes in sediment lithology and microbiota composition. The first cycle covers the intervals of 75 - 54 m in borehole 246 and 81 - 60 m in borehole 230. It thus corresponds, in the first case, to the lower and middle parts of the Kiev Formation, when retaining its boundaries as defined by field research (see the figure) and, in the second borehole, to the entire volume of this formation.

The lower part of the Kiev Formation (interval 75 - 64 m, borehole 246) is characterized by a transition from siltstones to clays. The composition of radiolarians restricts this part of the section to the interval of the *Axoprunum visendum* assemblage, whereas the diatoms correspond to the upper part of the *Coscinodiscus* aff. *tenerrimus* Zone. The interval of 71 - 84 m (borehole 230), containing the *Axoprunum visendum* assemblage, is composed of carbonate-clayey siltstones with radiolarians, benthic foraminifers of the *Pseudo-*

clavulina subbotinae–Hopkinsina bykovae ucrainica Zone, and nannoplankton of the *Discoaster bifax* Zone. We correlate the facial changes at the base of the Kiev rock sequence and the extensive development and variability of the carbonate and siliceous plankton and benthos with the beginning of the transgressive cycle. The intervals of 64 - 52 m (borehole 246) and 74 - 61 m (borehole 230) are homogeneously composed of clayey siltstones. The disappearance of the carbonate and the paucity of the siliceous organisms indicate the regressive character of the deposits.

The second transgressive cycle is more clearly traced in borehole 230, where at the base of the Derizovka Formation the sands are overlain by clayey siltstones. The interval of 56 - 48 m contains abundant diatoms of the *Coscinodiscus succinctus* Zone and radiolarians of the *Heterosestrum shabalkini* assemblage. The appearance of a rich assemblage of siliceous organisms, similar to that from the base of the Kiev Formation, coincides with the transition from sandy deposits to sandy clays. We consider this sequence as the regressive in the cycle. In borehole 246, the deposits containing the corresponding assemblages of radiolarians and diatoms were detected in a member of clayey siltstones (interval 51 - 44 m), almost similar in lithology to the underlying rocks. According to the field description of the borehole, this member forms the top of the Kiev Formation, and we correlate it with the lower part of the Derizovka Formation on the basis of the identical composition of their radiolarians and diatoms. Therefore, at the base of the second cycle, there are the same facial changes from sands to clayey siltstones. The exact position in the cycle and the age of the overlying sandy unit, lithologically referred to the Derizovskaya formation, cannot be definitely determined because of a lack of paleontologic remains. The unit is probably deposited above an erosion surface and corresponds to higher horizons in the section.

As a result of this analysis, we may state that the Kiev and Derizovka (Khar'kov) formations are independent cyclically deposited sequences. Both contain, as a constituent part of the radiolarian and diatom assemblages, abundant and evolutionarily inert neritic species, whose presence complicated the stratigraphic interpretation of the Kiev and Derizovka microfossils. Moreover, the ecology of the Derizovka assemblage is different in comparison to the Kiev microbiota. Its neritic species among diatoms are more diverse, and the number of boreal species among radiolarians grows notably.

The most important aspect for the correlation of both sequences was the identification of the zonal or index species of the inter-regional scale among nannoplankton, radiolarian, silicoflagellate, and diatom assemblages of the Kiev rock sequence and among the latter three assemblages from the Derizovka deposits. The species of a definite distribution interval makes the dating of both sequences more accurate. The changes in lithologic and faunal composition of the deposits were used to reveal the cyclic changes in the structure of the Kiev and Derizovka formations, and to correlate them with the upper Lutetian–early Bartonian and Bartonian (early Bartonian?) transgressive-regressive cycles.

The present research was supported by the Russian Foundation for Fundamental Research (project no. 93-05-8876).

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Zoogeography of the Late Eocene Basins of Western Eurasia Based on Bivalve Mollusks

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Received May 31, 1993

Abstract – Zoogeographic provinces of Priabonian basins in western Eurasia are distinguished using restored generic and specific compositions of the bivalve mollusks and analysis of the distribution areas of their most characteristic genera and species. The tropical region of the Tethys is divided into Indo-African and Mediterranean subregions. Northward, we delineated the north European subtropical zone containing more differentiated fauna, which allows us to divide it into provincial subgroups.

INTRODUCTION

The high eustatic level of the global ocean during the late Eocene time gave rise to widespread and freely-communicating shelf seas in western Eurasia. The Tethys ocean separating Africa from Eurasia was the largest basin of that time. Northward, there was a complicated system of epicontinental and marginal seas with a belt of islands related to the Alpine fold system, which formed a boundary zone between these seas and the Tethys (see Fig. 3). The vast eastern basins of the West Siberia, Turgai, and Turan plates separated the Asian and East European platforms. Western, central, and southern Europe existed in that time as groups of islands bounded by Tethyan seas to the south, by the Atlantic to the west, and by the Norway–Greenland Sea gulfs to the northwest, where they were separated from the Atlantic by the Faeroe–Greenland sill (Udintsev, 1984).

Despite the widespread climate cooling during the late Eocene, faunas of all the above basins remained thermophilic, i.e., tropical or subtropical. The Eocene mollusks of these basins seem to represent the richest and most thoroughly studied malacofauna in the world, although the available data is incomplete and requires revision. The taxonomic composition of mollusks from the basins under consideration strongly differs on both the generic and specific levels, and thus is very suitable for zoogeographic analysis.

MATERIALS AND METHODS

Using my own samples and the collections of R.L. Merklin, O.V. Amitrov, G.Kh. Salibaev, R.K. Makarova, V.L. Stefanskii, A.A. Veselov, I.A. Korobkov, P.M. Aslanyan, Yu. Karagyuleva, A. Rusu, and N. Meszaros, I studied the upper Eocene bivalve mollusks from the Black Sea coast, Ciscaucasia, Transcaucasia, northern Ustyurt, Aral coastal province, Turgai, Central Asia, southern Bulgaria, and Transylvania. It was then possible to compare complete lists of genera and species from these basins with published data on the paleobasins of Europe and north Africa, and thus to reveal general distribution patterns of shallow bivalve assemblages in the west Eurasian seas.

The species of the assemblages were first listed for each region (e.g., for the Mainz, Aquitanian, and other basins, or their large areas). Then the lists were compiled as a summary table. As an entire species list of the very rich fauna of the late Eocene would be too large, only generic and subgeneric levels were considered suitable for such a table. To classify forms, I used the slightly modified systematics of bivalve mollusks published in American Treatise ... (1969). Taxa in the table were divided into groups according to their geographic distribution: I – genera and subgenera occurring only in the North African coastal zone; II – taxa of the Tethyan coastal zone; etc. (see Table 3). When grouping, I took into account the taxa distribution in the next time interval.

Graphs were used to show the extent to which the taxa are diverse and thermophilic (Fig. 1). Since nearly all the late Eocene species are extinct, the thermophilic characteristics of mollusks were estimated according to their climatic zonation in present seas (Popov *et al.*, 1993, Table 5). Then I calculated the number of common taxa (species, subgenera, and genera) for each assemblage pair and twin Preston's coefficients (see Starobogatov, 1970, for calculation method) representing their differences (Tables 1, 2). The geographic location of species and genera, which are indicated by their most characteristic taxa, are depicted by corresponding schemes (Fig. 2). All this information was used to distinguish the most significant zoogeographic boundaries, the rank of which was estimated mainly by comparison with the mollusks' present distribution in the basins.

CHARACTERISTICS OF ASSEMBLAGES

Northern Africa. Recently revised and representative data on bivalve mollusks are available for the Pria-

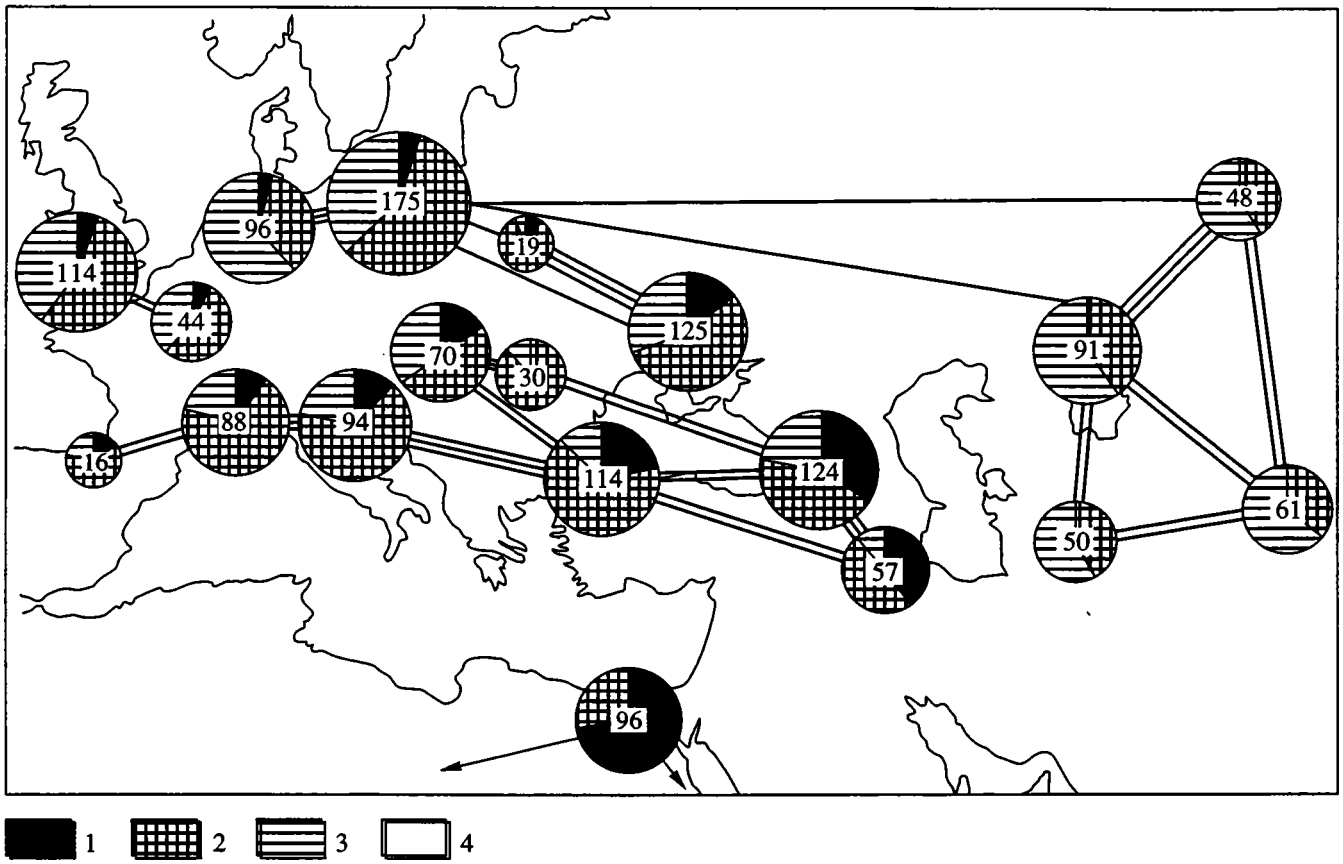

Fig. 1. Circle diagrams illustrating the relative abundance and climatic affinity of the late Eocene bivalve mollusks. Circle areas are proportional to the species numbers indicated in the center. Sectors indicate relative abundances of the following: 1 – tropical species; 2 – subtropical to tropical species; 3 – widespread species; and 4 – species mainly occurring in Boreal provinces. Connecting lines show the degree of compositional similarity between assemblages according to the threshold values of Preston's coefficients (see text for explanations).

bonian deposits in Egypt (upper part of the Mokattam Formation; Strougo, 1977). This fauna includes about 100 species of 70 genera and subgenera. Among them, 27 taxa are unknown for the Priabonian of Europe (Group I, Table 3), though some of them (*Porterius, Rostarca, Perna, Aviculoperna, Mimachlamys, Miodomeris*, and *Tugonia*) are found in the south, but mainly in the middle Eocene strata. Thus, our data on the Priabonian taxa in Europe are probably incomplete. Other genera (*Hawaiarca, Trisidos, Striostrea, Vasticardium, Quadrans, Notocorbula*, and *Carolia*) show that this fauna bears an affinity to the Indo-Pacific type. Some genera and subgenera were probably African endemics (*Protonoetia, Rectangularca, Cossmannella, Farisia, Crepispisula, Nicolbia*, and *Raetomya*). The last five genera are also classified as Eocene taxa from Algeria. The specific composition includes more than 90% of the species first described in north Africa, and 20 of them are common in Egypt, Somalia (Cabdulqadir and Carush, 1982), and India, whereas only 8 widespread species occur in southern Europe. A big difference between north African and European faunas is also evident in other benthic groups – the gastropods and echnoids.

Southern Europe. The most representative assemblages of mollusks have been described in the Priabonian stratotypes of northern Italy (Piccoli and Massari, 1968) and the Nummulitic Alps of southeastern France (Boussac, 1911). These assemblages consist of 80 - 90 species each, and of 59 and 44 genera and subspecies, respectively. About one third of the species and 32 genera and subgenera are common for both. About 70% of the species in both assemblages belong to tropical and subtropical zone genera, including *Pinna* s.s., *Trachycardium, Fimbria*, and *Discors*, which are characteristic of taxa of this type. Most species of the assemblages were typical of the Tethyan provinces, but many (30%) had a wider distribution range, and 19 species from northern Italy also occur in the middle Eocene (Bartonian) deposits of the Parisian basin.

In their generic composition, both assemblages are similar to the Eocene malacofauna from Macedonia, but the latter is poorer (36 species) and corresponds in age not only to the late Eocene, but also to its middle substage. The Priabonian mollusks from southern Bulgaria are rather representative (114 species of 63 genera and subgenera; Karagyuleva, 1964), though like others, this assemblage is far from complete. It lacks smaller forms, and fossils are represented mainly by large

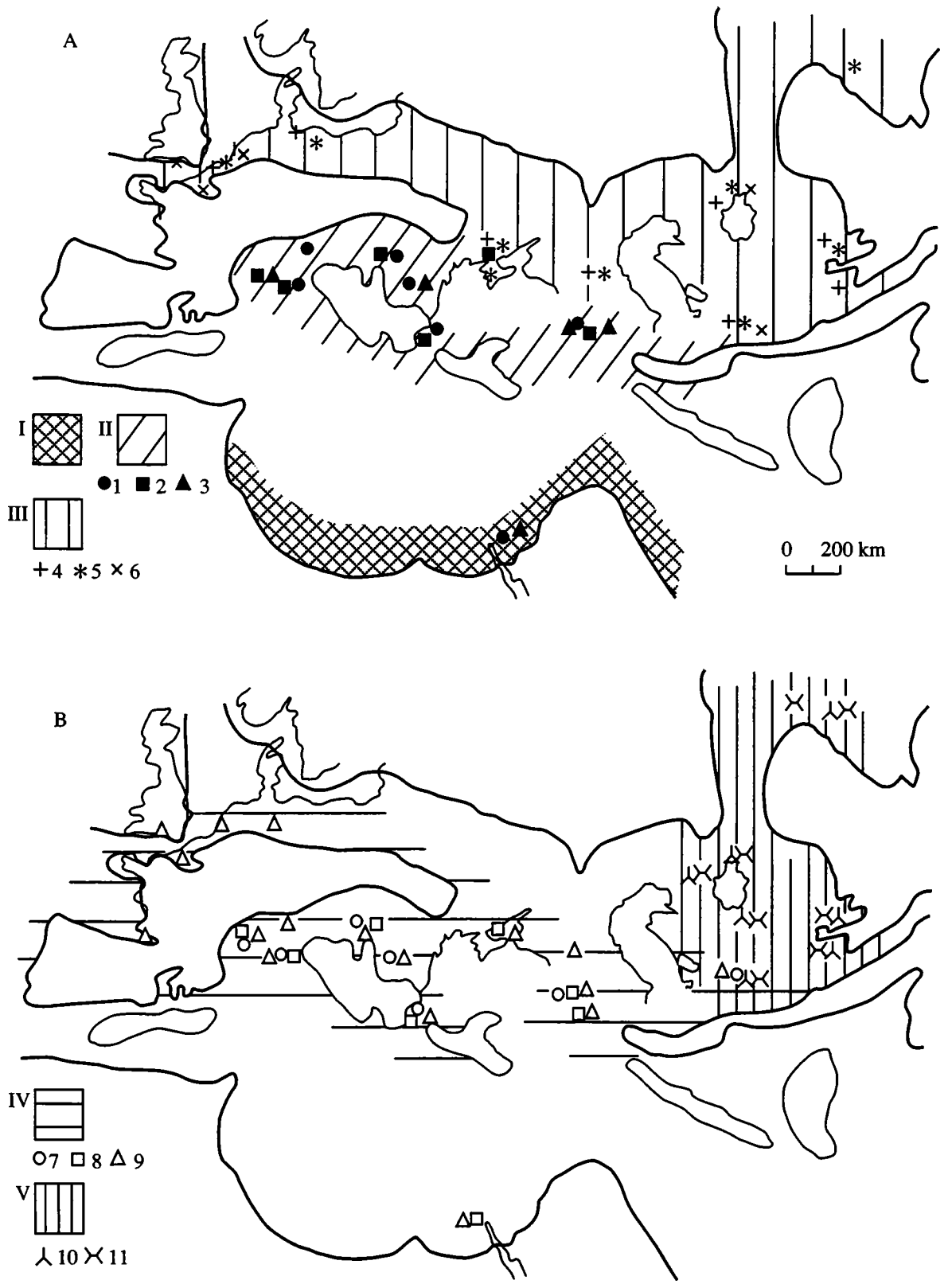

Fig. 2. Distribution areas of some characteristic genera and species of the late Eocene mollusks.

A. Distribution areas of the following: I – Indo-African genera; II – North-Tethyan genera: 1 – *Lima* s.s., 2 – *Crassatella* s.s., and 3 – *Macrosolen*; III – North European genera and subgenera: 4 – *Nuculana* (*Saccella*), 5 – *Astarte*, and 6 – *Lentidium*.

B. Distribution areas of the following: IV – widespread thermophilic genera: 7 – *Pecten* s.s., 8 – *Crassatella* s.l., and 9 – *Spondylus*; V – endemic Transcaspian species: 10 – *Nucula aralensis* and 11 – *Nemocardium razmyslovae*.

Table 1. Comparison of generic compositions of bivalve mollusks from the Priabonian deposits of Europe and North Africa

Preston's coefficients

Countries	1	2	3	4	5	6	7	8	9
(1) Egypt	**71**	*0.80*	*0.78*	*0.88*	*0.78*	*0.72*	*0.79*	*0.89*	*0.84*
(2) Southwestern France	13	**44**	0.43	0.62	0.54	0.69	0.50	*0.76*	*0.71*
(3) Northern Italy	18	32	**59**	0.57	0.40	0.59	0.55	*0.76*	0.63
(4) Hungary	10	21	27	**46**	0.52	0.76	0.65	*0.74*	*0.75*
(5) Southern Bulgaria	18	29	42	30	**63**	0.61	0.67	*0.73*	*0.64*
(6) Armenia	19	16	25	13	24	**44**	0.55	*0.92*	*0.71*
(7) Georgia	20	31	35	25	38	28	**70**	*0.71*	*0.62*
(8) England	8	14	17	17	19	6	21	**52**	*0.80*
(9) Southern Ukraine	17	21	30	19	32	20	34	17	**81**

Common genera — Number of genera

The diagonal row of values in bold shows the number of genera and subgenera in the assemblages under comparison; the amounts of common taxa in assemblage pairs are numbered below to the left, and Preston's coefficients for each pair, above to the right. The coefficients indicating a weak assemblage affinity are italicized, those for tightly related groups are in semibold.

thick-walled shells. *Macrosolen, Lima* s.s., *Trachycardium, Vasconella, Avicularium*, and *Corbula* (*Bicorbula*), which are characteristic genera and subgenera of this fauna, suggest its similarity to assemblages from southern Europe and Transcaucasia. At the same time, this fauna comprises many (40.33%) widespread species, which occurred outside the Tethys.

Table 2. Comparison of species compositions of bivalve mollusks from the Priabonian deposits of northern Europe and Transcaspian regions

Preston's coefficients

Countries	1	2	3	4	5
(1) Belgium	**96**	0.43	0.86	0.86	0.83
(2) Germany	75	**175**	0.85	0.86	0.78
(3) Kazakhstan	16	22	**88**	0.68	*0.90*
(4) Central Asia	20	24	40	**121**	*0.89*
(5) Southern Ukraine	23	39	13	18	**130**

Common genera — Number of genera

Transcaucasia. Fauna from the Rind Horizon of Armenia includes about 60 species of 44 genera and subgenera of bivalve mollusks. Some of its genera, such as *Pinna* s.s., *Lima* s.s., *Trachycardium, Avicularium*, and others, indicate that the fauna inhabited warm, probably tropical waters and is comparable to previous assemblages. The same is evident in its typical Tethyan forms: *Chlamys subdiscors, Spondylus cisalpinus, S. castellanensis, Crassatella vapincana, Trapezium alpinum*, and others.

In Georgia, most rich faunas of the late Eocene occur in the Akhaltsikh depression (76 species) and western Georgia. In total, these bivalve mollusks belong to more than 120 species of 70 genera and subgenera. The assemblage is north Mediterranean, as indicated by available specimens of *Glycymeris* (*Tucetilla*) *jacquoti, Chlamys biarritzensis, Ch. subdiscors, Spondylus cisalpinus, Macrosolen hollowaysii, Trapezium alpinum, Katelysia* (*Textivenus*) *texta*, and others. Still, despite the similar facies of enclosing rocks (coastal sandstones), the fauna contains only 15 species (13% of their total amount) common with the Rind fauna of Armenia. It is probably natural that Georgian fauna lacks most thermophilic genera (*Pinna* s.s., *Lima, Trachycardium*, and *Avicularium*) characteristic of the Armenian assemblages. In addition, this fauna is richer (44.37%) in widespread species.

Central Europe. The most complete data is available for the late Eocene faunas of Hungary and Rumania. The Hungarian assemblage of bivalve mollusks contains about 70 species of 46 genera and subgenera. Among typical north Mediterranean genera, it includes *Vasconella*, *Pinna*, and *Lima* s.s.; in addition, there are various forms of *Spondylus* and *Crassatella* genera and characteristic Tethyan species – *Glycymeris jacquoti, Modiolus incomptus, Chlamys biarritzensis, Spondylus cisalpinus*, and others. Also, many species of the assemblage had a wide geographic distribution. The mollusk assemblage from Rumania (Transylvania) has very similar features, and contains genera and species of north Mediterranean affinity (*Pinna, Lima, Vulsella dubia, Chlamys biarritzensis, Ch. subdiscors,* and others). The poor fauna from Austria and richer (51 species) assemblage of bivalve mollusks from Bavarian Alps (Schlosser, 1925) are the same.

Western Europe. The Headon, Osborne, and Bembridge beds in England and the Ludes Beds in the Parisian basin (Aubry, 1986) represent the Priabonian deposits of the region. The richer fauna from England includes more than 100 species of 52 genera and subgenera; a new revision would probably increase the number of the latter taxa. Nearly half of the species here was inherited from the Bartonian faunas, whereas the Priabonian mollusks were less diverse due to the elimination of the most thermophilic genera. Variable fauna forms include venerids (10 species), tellinids (8), lucinids (6), and corbulids (5). Characteristic Tethyan species are absent with the exception of widespread forms (*Arca biangula, Orthocardium porulosum, Panopea intermedia, Cubitostrea plicatula*, and others). The number of thermophilic genera is limited and includes *Arca, Barbatia, Pteria, Megaxinus* (*Gibbolucina*), and "*Crassatella*."

The Ludian deposits of the Parisian basin yielded 44 species of 33 genera. In thermophilic affinity and generic composition, the assemblage is similar to the fauna from England, though common forms are rare. The poor correlation between both assemblages is probably due to the different salinities of the relevant basins.

As British and Parisian assemblages are so different from the Mediterranean fauna, it is interesting to also consider the assemblage from southwestern France (Aquitania). The composition of Priabonian mollusks here is poor (16 species of 15 genera) but characteristic: 7 species were first described in Aquitania, 7 are common in the north Mediterranean assemblages, and 2 belong to widespread forms. Thus, if Cossmann was right in indicating that the middle Eocene (Biarritz

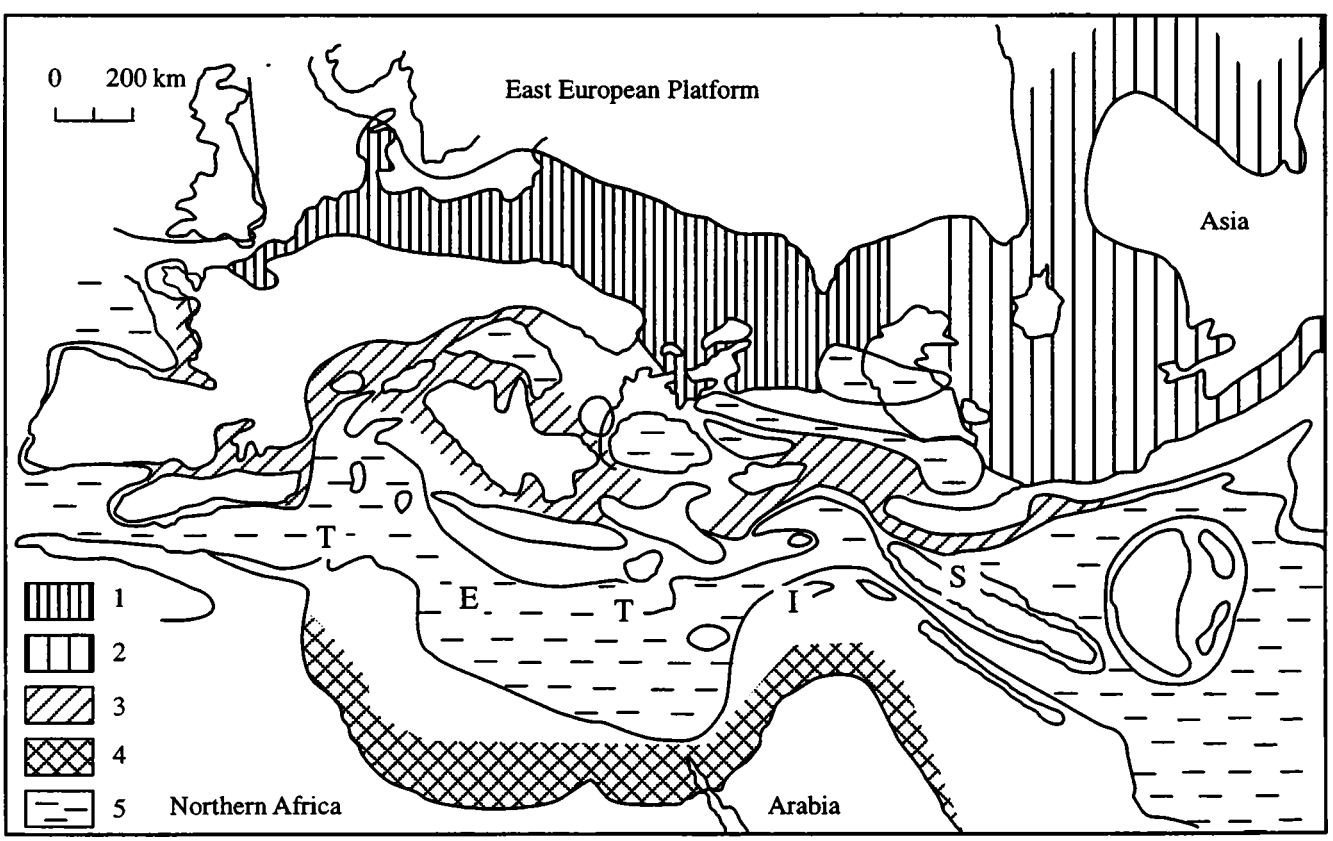

Fig. 3. Zoogeographic zonation of the West Eurasian basins based on the Priabonian bivalve mollusks.

1, 2 – North European region and 2 – its Turan province; 3, 4 – Tethyan region: 3 – subdivided into Mediterranean and 4 – Indo-African subregions; 5 – deep-water areas.

Table 3. Distribution of bivalve mollusk taxa in the late Eocene basins of western Eurasia (numbers indicate the amount of species in the taxon)

Genera and subgenera \ Localities	Egypt	Southwestern France (Alps)	Northern Italy	Hungary	Rumania	Southern Bulgaria	Armenia	Georgia	England	France	Southern Ukraine	Austria	Belgium	Germany	Kazakhstan	Central Asia
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
(I) Taxa occurring only in northern Africa																
Calorhadia	1															
Barbatia (*Rostarca*)	2															
B. (*Hawaiarca*)	1															
Trisidos	1															
Porterius	1															
Protonoetia	1															
Rectangularca	1															
Perna	1															
Aviculoperna	1															
Eburneopecten	1															
Chlamys (*Mimachlamys*)	2															
Carolia	1															
Ostrea (*Turkostrea*)	3															
Striostrea	1															
Nicaisolopha	2															
Cossmannella	2															
Farisia	1															
Miodomeris	1															
Paraglans	4															
Trachycardium (*Vasticardium*)	2															
Crepispisula	1															
Nilobia	1															
Quadrans	1															
Tugonia	1															
Corbula (*Notocorbula*)	1															
Raetomya	1															
(II) Taxa whose distribution is limited by the Tethys basin																
Vasconella		2		1		2		1								
Atreta					1											
Pinna s. s.	1		1	1	1		1									
Lima s. s.			3	1	1	1	1					3				
Lopha			2			1										
Eomiltha	2					3										
Divalinga	1							1								
Cardita s. s.							1									
Carditamera	?															

Table 3. (Contd.)

Genera and subgenera \ Localities	Egypt	Southwestern France (Alps)	Northern Italy	Hungary	Rumania	Southern Bulgaria	Armenia	Georgia	England	France	Southern Ukraine	Austria	Belgium	Germany	Kazakhstan	Central Asia
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
Glyptoactis	2	1														
Trachycardium	1		1			1	1									
Vepricardium	1						1									
Europicardium							1									
Avicularium			1			1	1									
Tellina s. s.							1	2								
Macrosolen	1	1			1	1		1								
Azorinus								1								
Katelysia (*Textivenus*)	1							2								
Mysia				1												
Corbula s. s.						3		2								
C. (*Bicorbula*)	3	1	2			1		1								
Pholadomya (*Procardia*)				1												
(III) Taxa occurring north off the Tethys basin in the Priabonian only																
Barbatia (*Obliquarca*)											1					
Cucullaria											1					
Trinacria									1							
Palliolum (*Lissochlamys*)											1					
Lucina s. s.	1		1			2	1	1			1					
Linga						1										
Megaxinus s. s.							1	?			1					
Myrtea								?								
Miltha			1				2	?		1						
Codalucina (*Epilucina*)											1					
Barbierella											1					
Pseudomiltha	1			1		2	2	3			1					?
Divaricella									2							
Boeuvia						1	1	1			1					
Fimbria			1		1	1	2				2					
Glans			1				2	1			1					
Venericardia		1				1	2	2			1			1		
Pteromeris											1					
Carditella											1					
Crassatella		6	1	2		4	1	3			1					
Cardium			2	1					3			1				
Granocardium	?															
Cardilia										1						

Table 3. (Contd.)

Genera and subgenera \ Localities	Egypt	Southwestern France (Alps)	Northern Italy	Hungary	Rumania	Southern Bulgaria	Armenia	Georgia	England	France	Southern Ukraine	Austria	Belgium	Germany	Kazakhstan	Central Asia
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
Asaphinella										1						
Periglypta	1															
Tivela				1												
Tivelina	1															
(IV) Widespread taxa																
"Nucula"		1		1		2			4	2				1		
Nucula s. s.								1			1					
N. (*Lamellinucula*)	1						1	1					1	2	5	4
Leionucula	1														1	
Acila (*Truncacila*)	1															
Nuculana						1			2							
Neilonella															1	
Arca	1		2	1		2		1	1	1	3			2		
Acar						3					1			2		
Barbatia s. s.	2	2	1	1	2	3		2	2	1	6	1		4	1	1
B. (*Plagiarca*)											1			5		1
Anadara					1						1		1	1		1
Bathyarca														2	1	1
Arcopsis											3			2		
Limopsis s. s.				1					1		1					
L. (*Pectunculina*)		1	2			1		1			2	1	3	2	1	1
Glycymeris s. s.	2			1		1					1	1	1	2		1
G. (*Tucetilla*)		1	1	1		1		2			3	1				
Mytilus						1		1	1		?					
Arcomytilus		1	1			?								1		
Crenella			?						1					2		
Arcoperna										2				1		1
Musculus	1				2			2	3				1	1	1	2
Modiolus		2	2	1	1	2	1	?		1			1	1	2	3
Lithophaga			2			1									1	
Atrina				1			1	1					1	1	2	2
Pteria				2		1		1	1	1		1	1		1	
Isognomon																1
Vulsella	2		2		1	1	1				3			3		
Lentipecten		1	1	1	1	1	1	1			1	1	1	1	1	1
Propeamussium s. s.				1		1						2				
P. (*Parvamussium*)		2	1			2		1				2				2

Table 3. (Contd.)

Genera and subgenera \ Localities	Egypt	Southwestern France (Alps)	Northern Italy	Hungary	Rumania	Southern Bulgaria	Armenia	Georgia	England	France	Southern Ukraine	Austria	Belgium	Germany	Kazakhstan	Central Asia
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
Palliolum (*Delectopecten*)													1		1	
"*Chlamys*"		2	1	3	1			1				2		3		7
Chlamys s. s.		2	1	1	1	2		3	1		2			1		
Ch. (*Aequipecten*)		1	1	1	1	5	1	2			3		3	2	2	3
Ch. (*Hilberia*)	?												1		?	
Pecten		1	4	3	1	1		1								1
P. (*Flabellipecten*)															1	
Plicatula	2		1								2			2		
Spondylus	1	4	5	9	3	5	4	5			4		1	2		
Dimya			2	1		1					2			2		
Anomia			3	2		2		1	1	2	2				1	1
Ctenoides						1					3			2	1	
Limatula											1			1		
Limea											1			1		
Ostrea			1	1		2			1	1			1	·		
Cubitostea	1	2	1			3		4	2		2			2	2	3
Crassostrea			2													
Pycnodonte		2	3	1	1	2	2	3				2	2	2	2	3
Amphidonte																1
Gryphaea (*Ferganea*)																1
"*Lucina*"		5		1				1	12	2		2				
Cavilucina s. s.						1										
Megaxinus (*Saxolucina*)		1	1		1			?			1	1				2
M. (*Gibbolucina*)	2					1	1	2	1		2			1		
Parvilucina								2			3		1			
Lucinoma								1								
Claibornites											1					
Ctena											1			1		1
Pegophysema													1			
Thyasira					1						1		1		2	4
Diplodonta	1						1		3	2			2	2	1	1
Felaniella	1															
Chama	2	2	3	1	2	1	1	2		1	4			1		
Erycina														2		
Kellia									1					1		
Pseudopythina													1			
Lepton									2				1	2		

Table 3. (Contd.)

Genera and subgenera \ Localities	Egypt	Southwestern France (Alps)	Northern Italy	Hungary	Rumania	Southern Bulgaria	Armenia	Georgia	England	France	Southern Ukraine	Austria	Belgium	Germany	Kazakhstan	Central Asia
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
Spaniorinus													1			
Sportella														1		
Anisodonta													2	1		
"Cardita"				2	1				4							
Acturellina		1	2			1	1	1		1	5		1	1		
Cardites			1			1	2	2			1	1	1		4	5
Carditopsis											2		1			
Sita													1			
"Crassatella"		2		3					1							
C. (Landinia)			1								2			1		
Crassatina	1		2	1		1				1	5		1	3		1
Eucrassatella		1					1	1		1	3		1	1	3	7
Laevicardium	2						1	2				1	1			
Loxocardium							1	1					1	1		
Fragum	?						1				1			2		
Plagiocardium								1	1	1	1					
Orthocardium								1	1	1	1		1	1		
Acanthocardia			2				2	2								
Nemocardium s. s.	1	1	1	2		1	2	4	2		1		1	2	1	1
N. (Habecardium)								3						2	2	1
Discors	1		1			1	1	2					1	2		
Goniocardium											1			2		
"Mactra"								?	2						1	
Lutraria															1	
Solen		1	1	1		1					1		1	1	1	2
Cultellus			1							1			1	1	3	3
Ensis									1							
Siliqua								1					1			
"Tellina"				1		3			8			1		1		
Angulus			1		1			1					3	5	1	1
Arcopagia	4	1	1				1	1					1	1		
Oudardia													1			
Macaliopsis													1	2		
Elliptotellina											1			1		
Arcopagiopsis											1			1		
Florimetis																1
Gari s. s.		1	1			1		1	3						1	4

Table 3. (Contd.)

Genera and subgenera \ Localities	Egypt	Southwestern France (Alps)	Northern Italy	Hungary	Rumania	Southern Bulgaria	Armenia	Georgia	England	France	Southern Ukraine	Austria	Belgium	Germany	Kazakhstan	Central Asia
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
G. (Gobreus)	1	1														
Psammobia			1	2						2				3		
Abra										3			2	3	1	
Solecurtus	1	1	1										1	1		
Congeria									1							
Arctica	1							2	1		1		1	1	1	1
Petalocardia									1		1		1	5		
Pygocardia								1						1		
Lutetia									1	1			1	2		
Trapezium		2	1			1	1	1			?				1	1
Coralliophaga											3		1	2		
Glossus								4	1		1		1			2
Miocardiopsis	1				1		1						1		1	2
Corbicula									1							
Polymesoda		4	1			1			5							
Spherium									2	1						
"Venus"			1													
Circe														1		
"Meretrix"		5	4	1		2		2	10	1	1			?		
Callista					1	4				1				2	1	2
C. (Chionella)											1					
Costacallista									1				1			
Chione			1			1										
Pitar	4												1	1	1	1
Pelecyora (Cordiopsis)		1	1		1	1			1		1			1	2	2
"Corbula"		4	1	1	1				4	3				1		
Corbula (Caryocorbula)	2										2			2	1	
C. (Ficusocorbula)								1		1					1	1
C. (Varicorbula)		1								1	1		1	2	1	1
Caestocorbula									1		1		1	1	1	1
Erodona									4							
Gastrochaena									1				1	1		
Eufistulans													1	1		
Hiatella													1	1		1
Panopea	1	3	1	1	1	2	1	4	1		2	1	1	1	2	3
Aspidopholas															1	
Martesia									1							

Table 3. (Contd.)

Genera and subgenera \ Localities	Egypt	Southwestern France (Alps)	Northern Italy	Hungary	Rumania	Southern Bulgaria	Armenia	Georgia	England	France	Southern Ukraine	Austria	Belgium	Germany	Kazakhstan	Central Asia
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
Jouannetia														1		
Teredo			3			1		1	1					2		
Pholadomya s. s.				3		1		1		1						
Ph. (*Bucardiomya*)		1	2				1	2						2	3	4
Thracia	1	4	1	1		2	2	3				1	1	2	3	2
Cuspidaria				2								1		5	1	1
Cardiomya													1			
Verticordia														1		
Pechiolia				1										1		
Clavagella s. s.	1									1					1	
C. (*Stirpulina*)								1			1		1	1	1	
(V) Taxa nonexistent in the Tethys basin																
Nuculana (*Saccella*)											3		3	7	2	3
N. (*Jupiteria*)													1		1	
Yoldia															1	1
Yoldiella															1	
Solemya												?				
Nucinella											1		1	1		
Nucunella											2			1		
Gregariella													1			
Gonimyrtea											1		1	1	1	1
Loripes													1			
Scalaricardita											3		1		1	1
Astarte											1		4	7	1	1
Digitaria														1		
Tridonta														2		
Parvicardium											1		1			
Sanguinolaria													1			
Aralocardia													1	1	2	3
Dosiniopsis												1				
Sphenia									4	2				1		
Lentidium									1	1			1		1	
Spheniopsis														3		
Cyrtodaria															2	2
Number of species	96	82	94	70	30	114	57	124	114	44	135	28	96	175	88	121
Number of genera and subgenera	71	44	59	46	25	63	44	70	52	33	81	20	76	96	59	58

time) faunas from southwestern France and the Parisian basin were similar, having no zoogeographic connections with the Mediterranean provinces, then in the Priabonian such connections became very probable and, as time went on, absolutely evident, as indicated by the richer Stampian fauna.

Northward, the representative mollusk assemblages of western Europe (from Belgium, Netherlands, and Germany) are probably of a younger, late Priabonian age. The Latdorfian assemblage of northern Germany is the richest among them and contains 175 species of bivalve mollusks (Koenen, 1893 - 1894). The taxonomic diversity of this fauna, which comprises abundant arcids and genera such as *Coralliophaga, Vulsella, Plicatula, Spondylus, Dimya, Ctena, Crassatella* (*Landinia*), *Discors*, and *Goniocardium*, suggests the thermophilic, or at least subtropical, character of this assemblage.

Many species (nearly half) described by Koenen are still considered endemic. Many other smaller forms of the genera *Sportella, Erycina, Kellia, Lepton, Anysodonta, Lutetia*, and some others are very unstable under fossilization, and thus poorly studied.

The assemblage of bivalve mollusks in the Belgium lower Tongres (from the Glimmertingen Sands) (Glibert and Heinzelin, 1954) is poorer than the Latdorfian (96 species of 76 genera and subgenera). In addition, it seems less thermophilic, having few arcids and lacking many warm-water Latdorfian genera. Despite this, both assemblages are very similar in composition. They include 75 common species, or 190 of 244, if both bivalves and gastropods are considered.

The poor assemblage from Netherlands is entirely within the compositional range of the Latdorfian fauna.

Eastern Europe. The Priabonian Mandrikovka assemblage of Ukraine is very rich (not less than 130 species of 81 genera and subgenera). It includes tropical and subtropical species of *Spondylus, Fimbria, Venericardia* s.s., *Fragum, Goniocardium*, and other genera. Despite the similarity in generic composition between this and the Mediterranean assemblages, the Ukrainian fauna is nearly barren of the characteristic Tethyan species. Its 18 species are similar to forms from Anglo-Parisian basin localities (16 of them are similar to the Bartonian forms), and the number of species similar to those of the Latdorfian age is 39. Such compositional peculiarities, intermediate between the Bartonian and Latdorfian, probably reflect the early Priabonian age of the assemblage (it corresponds to the nannoplankton zones NP19 to NP20).

In eastern Poland, 13 of the 19 known bivalve mollusk taxa have analogues in the Mandrikovka fauna, and 7, in the Latdorfian (together with gastropods, the latter and Ukrainian assemblages share 23 common forms).

Transcaspian region. The known bivalve mollusks of the region differ greatly from the above-mentioned assemblages in their impoverished composition on the generic and family levels. The characteristic Priabonian taxa of arcids, lucinids, venerids, cardiids, and crassatellids were very poor, and typical assemblages included forms of the *Pholadomya, Panopea, Aralocardia, Nucula, Pelecyora, Gari*, and *Chlamys* genera along with mytilids and oysters. Nearly half of the bivalve species were endemic Transcaspian forms.

The Chegan Formation of the northern Ustyurt and northern Aral region contains the most diverse mollusk assemblage (88 species). It has only 22 species similar to forms of the Latdorfian fauna, while 36 representatives (40%) are endemic, typical only of the Transcaspian region, and 20 of them occur only in the Chegan fauna. Many Transcaspian endemics are abundant, characterizing the assemblage type.

The Chegan Formation of Turgai yielded 48 bivalve mollusk species. The assemblage is similar to the previous one, with 3/4 of its forms being similar, although their host rocks were deposited in a different paleogeographic basin of the west Siberian lowland. The composition of mollusks suggests a more shallow basin type, where sea-water parameters deviated somewhat from the norm.

The Isfara-Khanabad and Sumsar beds of Central Asia, along with the main part of the upper Torymbeur Subformation of Kopetdag, correspond to the Chegan Formation, and yielded 120 species of bivalve mollusks, 40 of which are similar to the Chegan forms. The main forms are abundant Transcaspian endemics (*Nucula aralensis, N. turgaica, Nuculana khanabadensis, Nemocardium razmyslovae, Aralocardia eichwaldiana, Thracia asiatica*, and others), but there were also more widespread species, some similar to those found in the Latdorfian assemblage (*Nuculana crispata, Cubitostrea plicata, Corbula ficus*, and others). The species and genera similar to the Latdorfian taxa, but absent in the Chegan deposits, are noteworthy (*Cubitostrea cyathula, Pholadomya weissi, Caestocorbula henkeliusiana, Corbula conglobata, Anadara* cf. *sulcicosta*, and *Crassatina* cf. *woodi*). It is also important to note some of the thermophilic genera that are found here, but are unknown in the northern Transcaspian basins (*Barbatia* (*Plagiarca*), *Isognomon, Florimetis*, and larger lucinids), as well as a limited number of species occurring mainly in the Tethyan regions (*Pycnodonte brongniarti, Modiolus rutimeyeri*, and *Chlamys meissonieri*).

COMPARISON OF ASSEMBLAGES AND CHARACTERISTIC DISTRIBUTION AREAS OF GENERA AND SPECIES

The results of comparing the generic and subgeneric compositions of assemblage pairs are presented in Table 1 together with estimations of Preston's coefficients. The data obtained shows that faunas from southern and central Europe and from Transcaucasia are both interrelated and different from faunas of the north European and north African localities. In the case of a comparison of the species compositions, the difference would be even greater. The north African faunas are

nearly barren of species similar to the European assemblages. The dissimilarity between south and north European faunas becomes evident when tracing the distribution area characteristics of discrete species and genera. Among the species (and genera especially) of bivalve mollusks from western Eurasia, there were many widespread taxa occurring in the Priabonian faunas of both the southern and northern provinces of Europe (groups III and IV in Table 3). As for the most abundant forms, such distribution areas were typical of *Arca biangula, Barbatia appendiculata, Limopsis costulata, Chlamys bellicostatus, Lentipecten corneus, Megaxinus gibbosula, Fimbria lamellosa, Nemocardium tenuisulcatum, Panopea intermedia, P. woodwardi*, and other species.

At the same time, many of the genera and species that inhabited the Priabonian basins of southern to central Europe and Transcaucasia were absent in the northern provinces (Group II in Table 3). The most characteristic genera and species of this distribution type are *Vasconella* s.s., *Lima* s.s., *Trachycardium*, and *Corbula* (*Bicorbula*), most forms of *Crassatella* s.s., *Glycemiris* (*Tucetilla*) *jacquoti, Modiolus rutimeyeri, Pecten arcuatus, Chlamys biarritzensis, Spondylus bifrons, S. cisalpinus, Macrosolen hollowaysii, Trapezium alpinum*, and some others (Fig. 2).

It is possible to list more than 20 genera and subgenera that were rare or absent in southern faunas, but common in northern provinces (Group V in Table 3). Characteristic species of the northern regions were probably *Nuculana galeottiana, Anadara sulcicosta, Acar lamellosa, Acropsis pretiosa, Musculus nysti, Astarte kickxi, Corbula cuspidata*, and others. The northern faunas seem less interrelated than the southern assemblages and probably underwent a change in composition during the Priabonian time. For example, I believe that the peculiar composition of the Mandrikovka (South Ukraine) assemblage, which distinguishes it from the north European faunas, is time, rather than geographically, related.

The Transcaspian faunas differ from those of northern Europe in their impoverished composition at the generic level. Abundant endemic forms of these faunas distinguish them from other groups (Table 2).

ZOOGEOGRAPHIC ZONATION

The most evident difference in the compositions of Priabonian bivalve mollusks is between faunas of the southern (North Africa) and northern (Europe) coastal zones of the Tethys. Thus, the data on mollusks favors Douville's idea (Strougo, 1977), which distinguishes the Indo-African region as a high ranking zoogeographic unit. The above difference is greater than the 40% present incompatibility between the zoogeographic areas under consideration (Semenov, 1982). In contrast, data on nummulites shows a great similarity between the taxonomic compositions of their assemblages from southern Europe, India, and northern and eastern Africa (Strougo, 1977; Cabdulqadir and Carush, 1982). Taking this into account, one should probably consider the Indo-African basins as parts of a tropical subregion of the Tethyan region, where the deep areas of the Tethys and sublatitudinal currents were barriers to the migration of coastal mollusks (Fig. 3).

Less evident, but still recognizable differences exist between northern and southern types of Priabonian faunas from western Eurasia. Unfortunately, it is impossible to use data on recent faunas from the localities to interpret the obtained results. For instance, using considerable information on various invertebrate groups, Semenov (1982) estimated empirical difference indexes (Preston's coefficients) for zones of various zoogeographic ranks. But when comparing the suggested gradation with indexes estimated for fossil assemblages, one should refer every fauna to a separate province. Even for assemblages representing the same fauna type (e.g., the Latdorf and Tongres assemblages, or those from the Chegan deposits in the North Ustyurt and Aral localities), Preston's coefficients are near 0.5 (0.43 and 0.47 respectively), which is higher than values for modern provinces and corresponds to those used to distinguish superprovinces or subregions. To avoid such inapplicable results, I used the conditional index gradation, which corresponds more exactly to additional information and results in zoogeographic subdivisions comparable with modern distribution areas, as well as also climatic and other factors. The accepted values are as follows:

In comparison	Common fauna	Different subprovinces	Different provinces	Different regions or subregions
At the generic level	<0.40	0.40 - 0.49	0.50 - 0.69	>0.69
At the species level	<0.60	0.60 - 0.69	0.70 - 0.86	>0.86

Taking into account stable differences between south and north European mollusks (since the Mesozoic time) and data on other faunal groups, these values can be thought to discriminate the Tethyan (tropical) and North European (subtropical) zoogeographic regions. The latter is often identified in publications as the Boreal province (when applied either to the Tertiary or to older seas). This term is quite inconvenient, as it as related to climatic factors, while a zoogeographic subdivision must be determined only by fauna similarity or dissimilarity. Even when used to classify a paleoclimatic belt, the term should be compatible with available data on taxonomic diversity and thermal types of fauna from the present Boreal zone. Among the late Eocene faunas, there are no assemblages with proper characteristics, and even the Transcaspian taxonomic groups, the most monotonous in composition, correspond more closely to modern, impoverished subtropical assemblages than to those from the Boreal provinces.

The Tethyan region of the considered distribution areas included two subregions, the Indo-African and Mediterranean, in contrast to the unified North European region. The boundary between the main regions remains nearly constant, being defined by the number of faunal groups: bivalve mollusks, gastropods (Amitrov, 1993), planktonic foraminifera, nannoplankton, mammals, and floral assemblages. The northern boundary of the "nummulitic seas" of the Priabonian time gradually shifted southward during the Eocene. In the middle Eocene, large provinces of Poland and the Anglo-Parisian basin were occupied by nummulitic seas and inhabited by many mollusk taxa whose distribution was restricted, in time, only by the Tethys Ocean. During the Priabonian, this boundary separated the Anglo-Parisian and Aquitanian basins, and basins of the Alps and Carpathians had fauna of the Mediterranean type. The present territories of southern Bulgaria and Transcaucasia were also parts of the Mediterranean subregion. All these faunas show a high degree of similarity and should probably be considered a single superprovince.

Northward, there were basins with faunas of the north European type. Their assemblages were more differentiated, changing compositions throughout the Priabonian. The composition of mollusks in the Anglo-Parisian basin was peculiar, but it is difficult to clearly distinguish their provincial features here because of the euryhaline assemblages.

The more northern Latdorfian province included faunas of the North Sea basin. It is unclear whether the southern Ukraine was a separate faunal province, or the peculiar features of the Mandrikovka assemblage are, for the most part, time related.

The Transcaspian malacofaunas show a high degree of endemism and differences that are persistent through time. Here, Salibaev (1983) suggested distinguishing the Turan province by subdividing it into three subprovinces. On the contrary, I consider it divisible into two subprovinces: the northern one comprising Kazakhstan, Turgai and the southern part of the west Siberian basin, and the southern Tadzhik–Fergana subprovince also including the Kopetdag.

Reviewer M.A. Akhmet'ev

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Stratigraphy of the Oligocene–Miocene Boundary Beds in Primor'e

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Received June 9, 1993

Abstract – The upper Oligocene–lower Miocene boundary layers are described with data from a comprehensive study of ten sections located along the submeridional profile across Primor'e. The paleoclimate relevant to these layers corresponds to the period of cooling experienced by many regions of the North Pacific and recorded in the coeval micro- and macrofloras composed of representatives of Taxodiaceae, Pinaceae, Betulaceae, *Ulmus*, and *Salix*, among which thermophilic plants are scarce, while *Fagus* pollen and macrofossils are almost absent. Accordingly, we tentatively suggest naming this rock sequence "*Fagus*-free" until a resolution granting it the status of a formation or, perhaps, a horizon is adopted. The exact position of the Paleogene–Neogene boundary in these rocks is as yet undefined because of difficulties in correlating individual continental sections with the oceanic zonation scale. It is probable that the boundary coincides with the level between layers where SPA-2A to SPA-2B subassemblages of miospores are replaced. Possible reasons why the deposits are barren of *Fagus* remains and problems pertinent to their nomenclature are discussed.

INTRODUCTION

The Paleogene–Neogene boundary in Primor'e is drawn at the base of the Ust'-Davydovka Formation, and its stratotype is located on the Rechnoi Peninsula near Vladivostok (*Resheniya* ..., 1982). Still, not all geological and paleobotanical evidence corresponds to the official viewpoint, rendering it arbitrary. For example, the available imprints of *Trochodendroides* ex gr. *arctica* (Heer) Berry and other plants in the stratotype of the Ust'-Davydovka Formation are typical of the Paleogene in the southern Far East (Klimova, 1971), which contradicts the early-middle Miocene dating of this formation adopted at the end of the 1940s. The study of microflora revealed other discrepancies as well. The late Eocene–early Oligocene character of the spore-pollen assemblages of the Ust'-Davydovka Formation (Kundyshev, 1990), confirmed by our results, either disproves the correlation between the Paleogene–Neogene boundary and the base of this formation or indicates the Neogene age of some plants-edificators of the Paleogene floras, thus undermining the last criteria for the subdivision of the Tertiary deposits in Primor'e. These contradictions prompted us to reassess the problem of the Oligocene–Miocene boundary layers and to discuss two pertinent aspects: a) the stratigraphic aspect important in tracing the Paleogene–Neogene boundary from the geological survey, and b) the paleoclimatic aspect pertinent to the supposed cooling during that time period.

OBSERVATION DATA

To solve the above problems, we used the geological and paleobotanical data from ten sections covering the Primor'e from the Khasan area in the south to the Bikin River in the north (Fig. 1). Pavlyutkin carried out the study and integrated sampling of the sections. Petrenko, in collaboration with Vysochina, conducted the palynological research of 360 samples in order to determine the biostratigraphic subdivision and to correlate individual sections. The data on the leaf and seed remnants, supplied by Klimova, Dorofeev, and Nikitin, were incorporated, along with radiometric datings obtained by Ganzei by decay tracking.

DISCUSSION OF RESULTS

In Primor'e, the Tertiary rocks traditionally include two coal-bearing groups divided by a barren sequence, referred to as the Nadezhdinskaya Formation. The Paleogene age of the lower group (the Uglovoe, Maitun, Pos'et and Nazimovo formations) and the Nadezhdinskaya Formation has been unanimously accepted. The dating of the upper group (the Ust'-Davydovka Formation and its analog), however, is ambiguous, because the plant remains in some of its sections are of the Paleogene age, whereas in other sections, the enclosing rocks are definitely of the Neogene time. A comprehensive analysis of the geological and paleobotanical evidence leads us to the conclusion that the Ust'-Davydovka Formation (the Artem–Tavrichanka depression) contains Paleogene plants in the stratotype and is a part of the upper coal-bearing group, which means that its section is incomplete and lacks Neogene constituents.

This supposition is also prompted by sedimentary deposits in several depressions that are lithologically similar to the deposits observed in the Ust'-Davydovka stratotype, but bear Neogene flora. The nearest is the Pushkino depression, which, together with the Artem–

Tavrichanka depression, belongs to the same sedimentary basin (Fig. 1). The Tertiary deposits of that depression were studied in detail during hydrogeological prospecting. Their most complete section was recovered by Borehole 214 in the central part of the depression (Fig. 2).

The lower three recovered lithologic units are, according to the lithologic features of their rocks, analogs of the Uglovoe, Nadezhdinskaya, and Ust'-Davydovka formations located in the Rechnoi Peninsula. The floras of the first two units are similar to those in stratotypes; the flora of the third lithologic unit has its own peculiarities, and two characteristic layers were also distinguished in a stratotype section of the Ust'-Davydovka Formation. The lower layer is 80 m thick; it is composed of rhythmically alternating grey, brown-grey, and greenish-grey siltstones and mostly fine-grained sandstones with siderite macro- and microconcretions bearing remains of bivalve mollusks of the *Unio, Nodularia, Viviparus*, and *Cipangopaludina* genera, which were determined by Popova of the Limnological Institute, the Siberian Division of the USSR Academy of Sciences. The lithologic features of the layer resemble those of the lower (sandy-shale) subformation of the Ust'-Davydovka Formation, from which we collected leaf imprints determined by Klimova to be forms of *Cocculus* cf. *ezoensis* Tanai, *Quercus* cf. *kushiroensis* Tanai, and *Platanus* sp., which are usually the components of the Paleogene Ust'-Davydovka flora in the Rechnoi Peninsula (Klimova, 1983) and of the Oligocene floras on Hokkaido Island (Tanai, 1970).

The character of the spore-pollen spectra, representing one spore-pollen assemblage (SPA-I) due to their similarity, does not differ much from the standard spectra in the Ust'-Davydovka Formation stratotype. Their characteristic features are a moderate abundance of Betulaceae (3.4 - 4.8% except *Alnus*, which reaches anomalously high values in individual samples), and *Ulmus* (6.6 - 8.0%) and a diversity and abundance of pollen of the *Juglans, Fagus, Hamamelis*, and typical Paleogene taxons, such as *Engelhardtia quietus* (R. Pot.) Elsik., *Quercus gracilis* Boitz., *Q. graciliformis* Boitz., *Q. conferta* Boitz., *Q. forestdalensis* Tr., and *Liquidambar zaisanica* Pan. There are persistent, though single finds of pollen of the *Hamamelis, Corylopsis*, and *Streculia* genera and representatives of the formal genera *Tetraporites, Triatriopollenites*, and *Plicatopollis*.

The pollen assemblages of the underlying rocks, which are analogs of the Nadezhdinskaya and Uglovoe formations, share many similar features with SPA-1, though with fewer thermophilic and archaic Paleogene elements. In the southern Primor'e, the spectra composing SPA-1 are those of the lower Oligocene and Eocene–Oligocene transitional beds. The available data indicates that the lower layer containing SPA-1 is most likely the early Oligocene in age.

The upper unit is 90 m thick; its lithology is similar to that of the lignite subformation of the Ust'-Davydovka

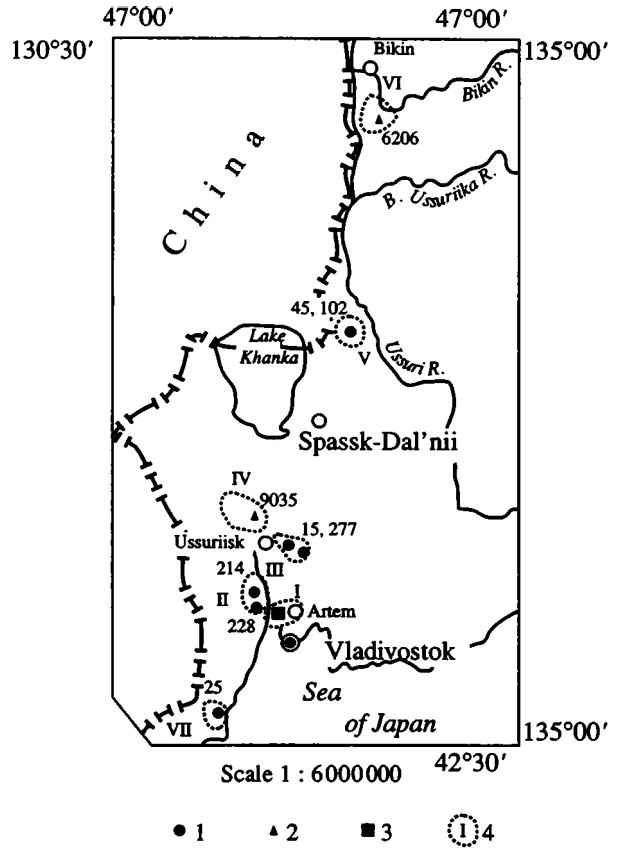

Fig. 1. Locations of the studied sections.

1 – Boreholes; 2 – quarries; 3 – stratotype of the Ust'-Davydovka Formation; 4 – depressions: I – Artem–Tavrichanka, II – Pushkino, III – Rakovka, IV – Pavlovka, V – Shmakovka, VI – the lower reaches of the Bikin River, VII – Poima.

Formation but differs in a small admixture of pyroclastic material in the silt fraction of the sediments and in having coal seams at the base. The paleobotanical differences are more significant. The impoverished assemblage of leaf flora from the upper bed (Borehole 214) contains only Betulaceae, *Salix*, and the remains of temperate *Ulmus* and *Juglans* genera not identified at the species level. The spore-pollen spectra also differ, and we consider them to be the SPA-2 assemblage. In its spectra, like in SPA-1, the angiosperms dominate, though not so obviously. The percentage of gymnosperms grows considerably owing to the presence of abundant pollen Pinaceae such as: *Pinus* s/g *Haploxylon*, *P.* s/g *Diploxylon*, and *Picea* (both s/g). In certain samples, amounts of the angiosperms and gymnosperms pollen are almost identical, and the number of spores is low. The Pinaceae reach their peaks of abundance when supplemented by dark conifers (up to 21%), among which *Picea* pollen, with a small admixture of *Tsuga* (1 - 2%), dominate. The abundance of Taxodiaceae pollen varies from 1.6 to 19.8%, but in most cases it is 5 - 8%.

The dominant angiosperms are forms of Betulaceae: *Betula* (up to 14%), *Carpinus* (up to 11%), *Corylus* (up to 10%), and *Alnus* (up to 14%); the *Juglans* and *Ulmus*

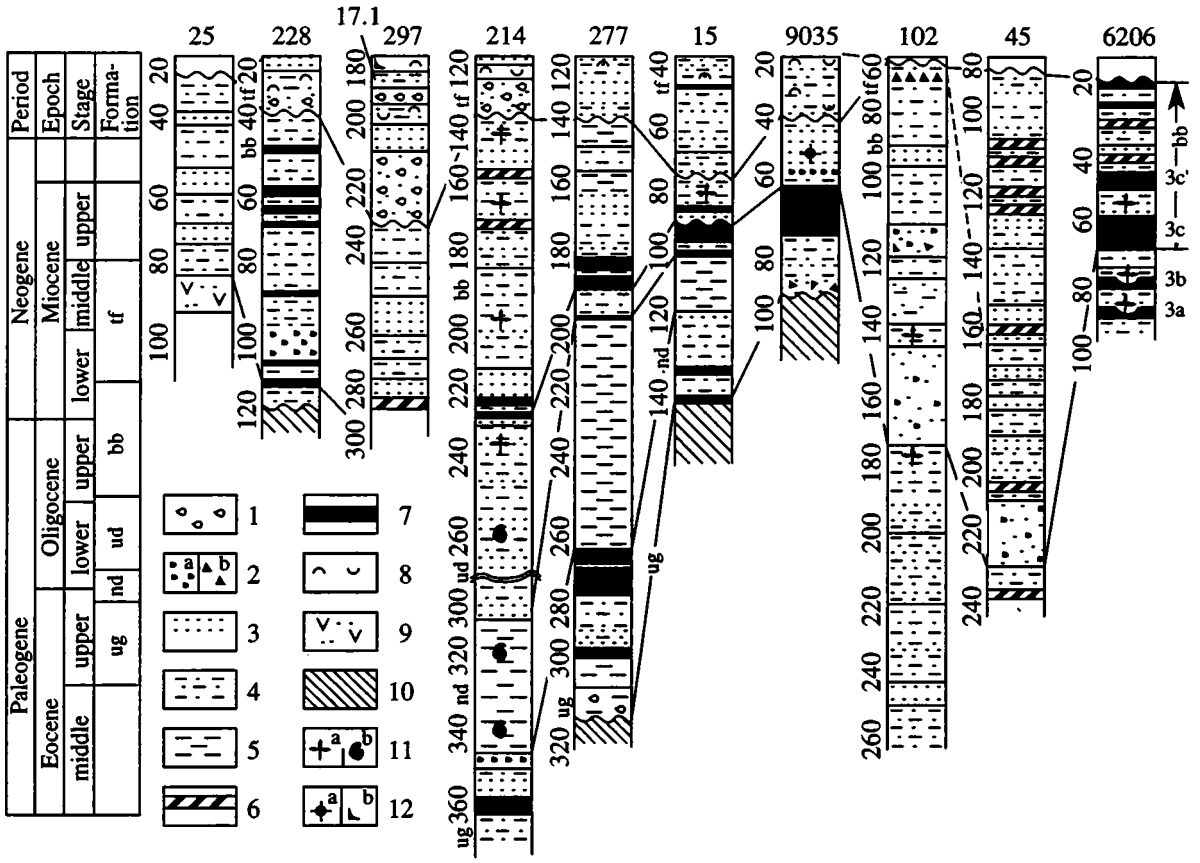

Fig. 2. Schematic correlation of studied sections.

1 – shingle; 2 – (a) gravel, (b) gruss; 3 – sand, sandstone; 4 – siltstone, silty sandstone; 5 – claystone; 6 – lignite, coaly siltstone; 7 – brown coal; 8 – tephrogenic material; 9 – andesite tuff; 10 – pre-Cenozoic rocks; 11 – (a) leaf flora, (b) fauna of mollusks; 12 – (a) carpolite flora, (b) radiological dating (Ma), ug – Uglovoe Formation; nd – Nadezhdinskaya Formation; ud – Ust'-Davydovka Formation; bb – *Fagus*-free Beds; tf – tephritoid beds.

groups lose their most thermophilic components, though the genera of temperate provinces still remain (*Juglans* 6 - 9% and *Ulmus* 14.1%). The spectra lack *Hamamelis* and *Corylopsis* pollen; the content of *Liquidambar* specimens decreases to single finds, and the pollen of the formal taxons is absent. The most striking feature of all eleven samples is a nearly complete lack of beech pollen (absent or single). Therefore, the upper bed (Borehole 214), lithologically similar to the lignite subformation of the Ust'-Davydovka Formation, essentially differs from it in the available plant remains, in general, characterized by Neogene features. It seems logical to refer the Paleogene–Neogene boundary in this section to the upper unit, which, for the sake of convenience, we shall temporarily call the "*Fagus*-free Beds."

The analysis of geological and paleobotanical data from the southern Pushkino depression showed that the *Fagus*-free Beds actually exist; this is not merely a coincidental combination of miospores in the pollen spectra. The beds can be clearly traced in boreholes and outcrops (Fig. 2). They overly the Paleogene or older deposits, and underlie the so-called tephritoid rocks with diverse lithology and pyroclastic material, which bear the Miocene palynoflora containing a considerable amount of beech pollen in all spectra. On the whole, these rocks have more thermophilic remains than the flora of the *Fagus*-free Beds. The tephritoid rocks were described in detail in an earlier paper (Pavlyutkin *et al.*, 1988). In the southern part of the Pushkino depression, the Miocene age of these rocks (17.1 Ma) was confirmed by radiometric dating of volcanic glass monofraction (Borehole 297, depth 197 m) by decay tracking at the Pacific Geological Institute, Far East Division, Russian Academy of Sciences (PGI FED RAS).

The pollen assemblages in all mentioned sections of the *Fagus*-free Beds in the Pushkino depression strikingly resemble the SPA-2 described above. Most of the data was obtained from Borehole 228. All 22 spore-pollen spectra from it are almost identical to SPA-2 (Borehole 214). The only difference is a certain increase in the content of dark conifers among the gymnosperms, mostly owing to *Tsuga* pollen (up to 8 - 9%) in combination with chestnut pollen (up to 4%) in some of the samples. This tendency was observed, however, in the uppermost rocks in the top sequence of Borehole 214. The pebbly rocks at the base of the *Fagus*-free Beds suggests erosion. The spore-pollen spectra and some

typical taxons refer the underlying flora to SPA-1, where the *Fagus* species dominate (18 - 57%), supplemented by the sclerophyllous evergreen (?) oaks *Quercus gracilis, Q. graciliformis*, and *Q. conferta*. These spectra contain abundant and diverse representatives of oaks and beeches. *Liquidambar* (3.9 - 26%), *Corylopsis*, and *Fothergilla* ("single") are some of the thermophilic components. The spectra of this type are present, according to our data, in the top rocks of the Ust'-Davydovka Formation located in the Rechnoi Peninsula.

The *Fagus*-free Beds were studied in detail in samples from boreholes 277 (int. 142 - 188 m) and 15 (int. 74 - 92 m) in the Rakovka depression (Fig. 2). The lithology of both sections is very similar to that of Borehole 214, but their rhythmic units, characterized by SPA-1 and bearing coal seams, have an insignificant thickness. There are no obvious indications of erosion at the base of the *Fagus*-free Beds in the cores of the above boreholes, though a hiatus probably occurred here, as implied by the thick coal layer (Borehole 277) or the bed of coarse-grained sand (Borehole 15) at this level.

The two sets of 16 and 7 samples of pollen spectra from *Fagus*-free Beds, recovered from boreholes 277 and 15, respectively, actually form one assemblage, with the taxonomic composition and percentage of components nearly identical to those of SPA-2 from Borehole 228. It also contains extremely rare beech representatives, i.e., not more than two pollen grains in individual samples. However, the chestnut pollen is noticeable (0 - 5.4%).

The core from Borehole 15 (int. 77 - 82 m) is poor in leaf flora represented by *Cercidiphyllum* cf. *crenatum* (Ung.) Brown, *Ulmus* cf. *carpinoides* Goepp., *U.* cf. *takayasui* Huzioka, *U.* cf. *quadrans* Goepp., *U.* cf. *protojaponica* Tanai et Onoe, *Zelkova* cf. *elongata* Suzuki, *Betula nathorstii* Suzuki, *B.* cf. *hommae* Huzioka et Nishida, *Carpinus* cf. *miocenica* Tanai, *Pterocarya* cf. *duensis* Fot.; *Juglandaceae* sp., *Populus* cf. *praemaximoviczii* Klim., *P.* cf. *balsamoides* Goepp., *Populus* sp., and *Salix* cf. *varians* Goepp. (after Klimova).

The listed plants constitute a part of the Miocene temperate and warm-temperate floras, but most of them cover wide stratigraphic ranges. They appear in the floras of Primor'e in the late Oligocene–earliest Miocene, thus indicating the transitional late Oligocene–early Miocene age of the rocks enclosing the *Fagus*-free Beds; at any rate, their presence does not contradict this supposition.

The analogs of the *Fagus*-free Beds were studied in the northeastern sector of the exploration trench (Exposure 9035) through a group of depressions of the Pavlovka brown-coal field. In this locality, the beds are mostly sandy and only a few dozen meters thick. They contain coal in various quantities and pyroclastic admixtures. The beds overlie the Paleogene coal-bearing deposits with a distinct contact, implying a stratigraphic hiatus, and are overlain either by shingles of the Suifun Formation (the Pliocene) or by tephritoid rocks. The latter contain Miocene macro- and microfloras with a considerable (up to 5 - 8%) amount of beech pollen in the spectra throughout and imprints of *Fagus* cf. *stuxbergii* (Nath.) Tanai and *Quercus* cf. *kazanskyi* Krysht. The pertinent radiometric ages are within the interval from 15 to 21 Ma (Pavlyutkin *et al.*, 1988).

The pollen assemblage from the layers underlying the *Fagus*-free coaly deposits definitely has Paleogene features and is somewhat similar to that of the Ust'-Davydovka Formation stratotype or SPA-1. The abundance and diversity of *Fagus* group pollen (oak, beech) associated with small tricoplate pollen of *Quercus gracilis* ("single"), *Q. graciliformis* (0 - 4%), *Q. conferta* (3.7 - 9.3%), and *Quercites sparsus* (Mart.) em. Samoil. ("single") distinguish the assemblage. Pollen of Hamamelidacea (*Fothergilla, Corylopsis, Hamamelis, Sycopsis*), *Altingiaceae* (*Liquidambar* – 2.7 - 9.7%), and sporadic *Loranthus* are also present, as well as pollen of the formal genera *Triatriopollenites, Triporopollenites, Rhoipites*, and *Verrutricolpites*.

The spectra of the *Fagus*-free Beds themselves retains all the features of SPA-2 (Borehole 228). Beech and oak pollen occurs in them as single grains; chestnut pollen is sometimes noted (0 - 8.4%). The dark conifers are unevenly distributed in the section, e.g., in some samples their content reaches 19% (including several species of *Tsuga*), in others it falls to 4 - 5%.

From sandstone strata (Exposure 9035) of the *Fagus*-free Beds rich in organic detritus, we macerated carpolite remains, in which P.I. Dorofeev of the Botanical Institute of the Russian Academy of Sciences identified *Epipremnum ussuriense* Dorof. sp. nov., *Betula* ex gr. *longisquamosa* Maedl., *Stewartia* sp., *Ampelopsis* sp., *Daphne* sp., *Pterocarya* sp., *Saururus* sp., and *Swida* cf. stolonifera Michx. This article does not aim to characterize each taxon in detail, and therefore, we note only that according to Dorofeev, they indicate the early Miocene–late Oligocene age of the flora. Later, we collected additional samples from this exposure and, according to Nikitin's determinations (Novosibirsk), their flora contains abundant remains of coniferous needles (*Pinus* sp., *Picea* sp., *Tsuga* sp.), seeds of *Chamaecyparis*, which are similar but not identical to *Ch. uralensis* Dorof. from the lower Miocene of the Middle Urals, *Epipremnum* ex gr. *cristatum* Nikit., *Aracispermum* ex gr. *canaliculatum* Nikit., *Comptonia* sp., *Pterocarya* sp., *Ostrya* sp., *Aldrovanda* ap., *Crataegus* sp., *Euphordia* sp., *Tilia* sp., *Stewartia* ex gr. *tavdensis* Mihi, and *Styrax* ex gr. *dravertii* Dorof.

This group of plants indicates, according to Nikitin, that the enclosing rocks are of the early Miocene–late Oligocene age interval, preferably of the early Miocene.

The analogs of the *Fagus*-free Beds were also studied in the cores from boreholes 45 and 102 drilled in the western Shmakovka basin, in the area of the Khanka depressions, whose Paleogene–Neogene deposits are poorly studied. The sections recovered from the holes

supplement one another. The *Fagus*-free Beds overlie a rather monotonous sequence of greenish-grey siltstones interbedding with thin fine-grained sandstones and lignites containing inclusions of sideritic microdruses. These rocks yielded pollen spectra (12 samples) of similar composition, which were identified as a spore-pollen assemblage of obviously Paleogene character according to the presence of such typical elements as *Gleichenia, Cyclocarya, Engelhardtia quietus, Quercus gracilis, Q. graciliformis, Q. conferta, Q. forestdalensis, Castanopsis, Fothergilla, Corylopsis, Hamamelis, Disanthus, Platanus, Rhus ustjurtensis* Boitz., *Loranthus elegans* Kulkova, and *Triartiopollenites, Triporopollenites.* All of the enumerated taxons are present in the spectra as "absent or single" grains. Based on the prevalence of *Fagus* and *Juglans* and a fair amount of *Liquidambar* (3 - 9%), we suggest that this pollen assemblage is, without a doubt, similar to the assemblage from the lignite subformation of the Ust'-Davydovka Formation in the Rechnoi Peninsula locality and SPA-1 from Borehole 214.

The lithology of *Fagus*-free Beds in the Shmakovka depression is noticeably different from that in the southern regions of Primor'e. They lack coal beds and contain only lignite interbeds in their lower parts. The amount of psephitic material, instead, forms as admixtures in the fine-grained (clayey-silty) rocks and grow high enough to form independent gruss interbeds, especially in the lower part of the section. Their gravel-size fragments mainly consist of the desintegration and rewashing products of the weathering crusts of the Khanka massiff. The rocks also contain less siderite, but all of the 92 spore-pollen spectra discovered indicate that the enclosing rocks actually belong to the *Fagus*-free Beds. Two subassemblages, analogous to those found in the sections of southern Primor'e and described earlier, were recognized as a result of the complete section and detailed sampling. The subassemblage SPA-2A (Borehole 102, int. 62 - 195 m; Borehole 45, int. 175 - 230 m) is identical to SPA-2 of the *Fagus*-free Beds in Borehole 214. Beech and oak pollen similar to the Mongolian species is present only in some samples as "single" grains, whereas the chestnut pollen is usually absent in the spectra.

The subassemblage SPA-2B (Borehole 45, int. 88 - 175 m) greatly resembles that of SPA-2 from the boreholes 15 and 277 and Exposure 9035. Oak and beech pollen is present as single grains in some spectra, but almost every sample contains chestnut pollen (up to 11%). This implies that the stratigraphic position of pollen spectra containing chestnut pollen is probably the uppermost in the *Fagus*-free Bed sequence.

A collection of leaf flora obtained from the lowermost part of the *Fagus*-free Beds (Borehole 102) includes, according to Klimova, *Tsuga* sp., *Sequoia* cf. *affinis* Lesq., *S.* cf. *langsdorfii* (Br.) Heer, *Metasequoia occidentalis* (Newb.) Chaney, *Glyptostrobus europaeus* (Br.) Heer, *Cercidiphyllum* sp., *Liquidambar* sp., *Ulmus* cf. *drepanodonta* Grub., *Celtis* cf. *nordenskioldii* Nath., *Castanea miomollissima* Hu et Chaney, *Alnus* cf. *schmalhausenii* Grub., *Betula* sp., *Cyclocarya* cf. *cycloptera* (Schlecht.) Iljinsk., *Pterocarya* cf. *asymmetrosa* Konno, *Juglans* cf. *miochinensis* (Tanai et Onoe) Suzuki, and *Alangium* cf. *aequalifolium* (Goepp.) Krysht. et Bors.

This floral assemblage contains early Miocene species (*Juglans miochinensis*, and *Acer megasamarum*), forms of a wider age range from the late Oligocene to Miocene (*Populus balsamoides, Ulmus drepanodonta*, and *Alnus schmalhausenii*), and Paleogene plants (*Cyclocarya* and *Alangium* cf. *aequalifolium*). Only the remains of *Castanea miomollissima*, common in the late Oligocene–middle Miocene floras of the Far East, belong to the *Fagus* group. On the whole, Klimova's assemblage indicates the transitional Oligocene–early Miocene time, probably the beginning of the Miocene.

In the northern Primor'e, the analogs of the *Fagus*-free Beds compose the uppermost part of the coaly sequence (Outcrop 6206) in the depression in the lower reaches of the Bikin River. The unit consists of three members represented (from the top downward) by greenish-grey siltstones, fine-grained sandstones, and several coal seams. The rocks contain numerous lignite laminae, sideritic concretions, and remains of freshwater mollusks. The unit base lies approximately at the bottom of Bed 3c (the uppermost in the third member). There are no indications of a significant hiatus in the contact zone apart from the pause in the abiogenic sedimentation related to the deposition of the 13m-thick Bed 3c.

The lithology of the underlying deposits is identical to that of the *Fagus*-free Beds. The obtained spore and pollen spectra include (as "absent or single" grains) typical Paleogene elements: *Dacrydium, Engelhardtia quietus, Quercus graciliformis, Q. conferta, Quercites sparsus, Q. Castanopsis, Hamamelis, Corylopsis, Disanthus, Platanus, Pseudoplicapollis, Rhoipites,* and *Verrutricolpites*. Spectra of similar composition with diverse and abundant *Fagus* (several species) and *Juglans* forms, but moderate amounts of birch and *Ulmus* pollen are typical of the lignite subformation of the Ust'-Davydovka Formation in the Rechnoi Peninsula.

The section of the *Fagus*-free Beds in the depression in the lower reaches of the Bikin River seems incomplete, lacking layers with SPA-2A and SPA-2B subassemblages. Its palinoflora is peculiar, because amid 18 analyzed samples, tentatively combined as SPA-2C, only some contain single grains of beech and none contain chestnut pollen. At the same time, they are rich in components that later formed the dark-coniferous taiga, i.e., *Tsuga* (several species) (1.5 - 16.4%), *Picea* (the total for both sections is 2.1 - 35.3%), and pine pollen (1.5 - 30%). Therefore, in the majority of spectra, the pollen of gymnosperms consistently dominates and that of *Ulmus* and birches is reduced. This is a principal feature of the assemblage.

So far, this palynoflora cannot be correlated with the other two palynofloras (SPA-2A and SPA-2B) because of the lack of continuous sections where we can trace

their relations to floras in the overlying rocks. Moreover, according to the gymnosperm abundance, this flora can sometimes be correlated with the upper Miocene palynofloras (Varnavskii *et al.*, 1988), but the geological evidence contradicts such speculations. The upper Miocene deposits (the Ust'-Suifun Formation) in western Primor'e are composed entirely of shingle deposits completely lacking coal occurrences. Their palynospectra, unlike in the *Fagus*-free Beds, contain noticeable amounts of beech pollen in almost every sample (up to 7.5%). The data on leaf flora from a lamina between beds 3c and 3c' (Kundyshev and Verkhovskaya, 1989) disputes the above speculation, because it completely lacks imprints of beeches, mainly composed of birch remains of species *Corylus jarmolenkoi* Grub, forming leaf peat, *Metasequoia disticha* (Heer) Miki, and rarely occurring *Alnus hokkaidoensis* Tanai, *Carpinus kushiroensis* Tanai, and *Ostrya antiqua* Grub. As Kundyshev correctly commented, these plants are common in the late Oligocene and Oligocene–early Miocene floras of Ashutasa and Hokkaido Island.

We can presume, of course, that the peculiarity characterizing the SPA-2C subassemblage resulted from the more northern location of the Bikin River section, but firstly, such a contrasting latitudinal zonation was unlikely to have occurred during the late Oligocene–early Miocene time, and secondly, the same type of spectra was distinguished in the uppermost part of coaly deposits of the Kraskino depression in the southernmost part of Primor'e (Varnavskii *et al.*, 1988). To the north, in the Poima depression, we studied a section of *Fagus*-free Beds composed of siltstones and fine-grained sandstones bearing scarce plant detritus (60 m). The beds overlie psephitic tuffs and andesite-basalt tuff breccias and underlie the Quaternary deposits. The composition of the pollen assemblage from 34 samples most closely resembles that of SPA-2C, though it contains fewer amounts of tsuga pollen of great species diversity.

The higher abundance of gymnosperms in SPA-2C, mostly at the expense of dark-coniferous components, indicates, in our opinion, not a different stratigraphic position, but rather the peculiar topography of the depositional settings. The depressions that yielded such pollen assemblages are bordered by fairly high (up to 1000 m) mountain ranges – the Chernye Gory and Strel'nikov Range – while others are located in plain areas with hummocky topography. Logical reasoning suggests that in the late Oligocene–early Miocene, the range slopes were covered with coniferous forests producing a mass of light pollen, which was buried in sediments of swampy lowlands. Further research will clarify this issue.

At present, geological and paleobotanical data allow us to consider the *Fagus*-free Beds as transitional upper Oligocene–lower Miocene deposits of the Primor'e. As the correct position of the Paleogene–Neogene boundary remains unclear because of difficulties in correlating continental sections with the oceanic zonation scale, it would be logical to date the *Fagus*-free Beds as late Oligocene–early Miocene, assuming that the Paleogene–Neogene boundary is located between beds with SPA-2A and SPA-2B.

The temperate *Fagus*-free flora correlates with the cooling event between the Oligocene and Miocene that, according to Sinitsyn (1980) and Fot'yanova (1988), occurred in the North Pacific and was according to Akhmet'ev (1976), of a global significance. This climatic minimum is reflected in the paleoclimatic curves based on isotopic ratios measured in shell remains of marine organisms from modern and ancient seas (Yasamanov, 1972; Shackleton and Kennett, 1975; Buchardt, 1978).

In Japan, the Aniai flora from the pertinent formation is the apparent analog of the *Fagus*-free flora. Its age which was earlier estimated to correspond to the second half of the early Miocene (Huzioka, 1964), is now 22 to 26 Ma (Kano and Yanagisawa, 1989), based on radiometric datings of the stratotype deposits of the Oga Island. In Sakhalin, the Hoindzho Formation, dated as the late Oligocene–early Miocene, contains temperate flora (Akhmet'ev, 1976a; Krasilov and Kundyshev, 1982; Fot'yanova, 1988); in Kamchatka, the Utkholok Formation of the same age contains ice-rafted pebbles at its base (Gladenkov and Shantser, 1989). In northeast Asia, the climatic minimum at the Oligocene–Miocene boundary is the most evident. The evidence from the reference section at the mouth of the Omolon River shows that flora of the Oligocene–Miocene boundary beds contain almost no broad-leaved plants and Taxodiaceae (Laukhin *et al.*, 1992). This type of forest appeared as early as the late Oligocene and remained dominant for the first 4 - 5 Ma of the early Miocene. In Alaska, the stratigraphic interval under discussion most likely corresponds to the Healy Creek Formation, whose flora bears almost no thermophilic elements (Wolfe and Tanai, 1980).

When discussing the *Fagus*-free Beds, it is important to analyze why there are no beech pollen and imprints in analogous floras because of the ambiguity in paleoclimatication of interpret *Fagus*-bearing beds as peculiar antipodes of *Fagus*-free Beds. For example, some researchers (Laukhin *et al.*, 1992; Akhmet'ev *et al.*, 1984) treat the abundance of beeches in fossil floras as decisive evidence of climate warming, while others (Krasilov and Alekseenko, 1977) consider this as an indication of cooling and beech forest migration to lower latitudes.

We believe that the relation between beech components in fossil floras and the paleoclimate is more complicated. In our case, a decrease in temperature was insufficient to cause the disappearance of beech from plant groups on plains and adjacent slopes. In spite of the late Oligocene cooling, the average temperature in January did not fall below +5°C during that period in Primor'e (Sinitsyn, 1980), whereas the corresponding critical value for beeches is known to be –6°C (*Zhizn' Rastenii*, 1980); and they normally develop under the January isotherm with a value of 0°C. For example, in Northern Honshu, there are widespread beech forests

with abundant *Fagus crenata* (Bukshtynov *et al.*, 1981). The same situation applies to the Carpathians, Poland, and Scandinavia, where beech forests with *Fagus sylvatica* are an important element of vegetation on the mountain slopes and in the plains (Alekhin, 1950). The lack of beeches in the late Oligocene floras of Primor'e results from the negative reaction of this group to the long periods of very dry air in the annual cycle. Modern geographic areas favoring beech vegetation have regular rainfall in all the seasons of the year (*Zhizn' Rastenii*, 1980) that are controlled by the global pattern of atmospheric circulation.

Analyzing the evolution weathering types on the territory of China, Wang (1984) concluded that, at the Eocene–Oligocene boundary, the entire system of atmospheric circulation in Asia underwent a drastic change under the effect of the Tethys decline, the Chegan Sea regression in western Siberia, and the uplift of the Tibet plateau. A similar explanation was advanced by Laukhin (1989). The new climatic system finally formed by the beginning of the late Oligocene, taking the form of the well-known pattern of monsoon circulation, which is the main modern climate factor in Primor'e.

Therefore, in winter, the long-term influence of extremely dry air masses from Central Asia had a drastic effect on the growth of beech forests. It is likely that the latter did not disappear in Primor'e during that time, but migrated, perhaps, to the upper mountain zones. In this case, the beech pollen, which is quite resistant to deflation (*Zhizn' Rastenii*, 1980), could not have reached the depositional settings in the lowlands. The macroremains of beeches had an even smaller chance of burial. We should emphasize here that north China is greatly affected by the winter anticyclone, and that beech forests do not grow north of the Qin Ling Range (Wang Chiwu, 1961; Kabanov, 1962), though the general temperature background and the average temperatures of January favor them, at least in the southern part of the region. For example, the more thermophilic (the minimal isotherm of January is –1°C) but more euryxeric chestnut is a common component of the forests on the northern slope of the Qin Ling (Bukshtynov *et al.*, 1981).

The presence of beech in the late Oligocene–early Miocene forests in north Japan, as testified by finds in the fossil floras of the Aniai type, was conditioned by the moderating effect of the powerful North Pacific warm current known as Kuroshio in the Japanese sector. However, there is the problem of how beeches restored their position in Primor'e in the later Neogene (late Pliocene including) epoches. We should remember that beeches are usual components in floras of the tephritoid deposits as both macro- and microfossils. The regeneration of beech forests (or the lowering of their border line) was likely related to the influence of the Sea of Japan on North Korea and Primor'e, for its opening occurred either at the early–middle Miocene boundary (Tsuchi, 1990), or somewhat earlier (Iijima and Tada, 1991). In the continental regions, e.g., in the Transbaikal region, the disappearance of beeches from vegetation at the end of the Oligocene was not replenished in later epoches, as is evident from the Neogene palynofloras (Kornutova, 1984).

In conclusion, we have to discuss some problems of nomenclature. The regional distribution, the considerable thicknesses, and also specific features of lithology and flora provide us with sufficient grounds to classify the *Fagus*-free Beds as a formation or horizon of perhaps super-regional significance, covering the stratigraphic range of the late Oligocene up to the first half of the early Miocene, inclusive. But then we have the problem of the so-called Pavlovka Formation. The Stratigraphic Glossary (*Stratigraficheskii Slovar'*, 1982) designates Gromov (Gromov and Gromova, 1960) as the author who identified and named this formation. But the cited work deals with almost the entire sequence of the Tertiary deposits of the Khankaiskii massif, which are 700 - 800 m thick. Moreover, Gromov never used the term Pavlovka Formation either in this paper or in any other of his publications. It was not mentioned in the summarizing review on geology of Primor'e (*Geologiya SSSR* ..., 1969). This name suddenly appeared in the stratigraphic scale, adopted at the Third Stratigraphic Meeting (*Resheniya* ..., 1982), but without any comments in the note to the resolutions. As a consequence, we face an abnormal situation, in which the stratigraphic scale unit of the formation rank has neither a clear authorship nor a concrete or, at least areal reference section, but yet represents a sedimentary sequence of considerable (for subplatform condition) thickness and has a "flexible" age range. Thus, the Pavlovka Formation is a fairly vague notion to be included into the regional stratigraphic scale of Primor'e as a definitive unit.

The name Pavlovka Formation cannot be used without breaking Rule IX.12, par. 2, of the Stratigraphic Code of the USSR (1977) when applied to the *Fagus*-free Beds, because they are only a part of the Tertiary sequence defined as the above formation in the Stratigraphic Scale of Primor'e. We do not suggest classifying *Fagus*-free Beds at the formation rank, because this requires, as a minimum, a selection of a stratotype section and its detailed characteristics, which demand special research. We only wish to emphasize the importance of studying these beds in terms of the International Project 326 "Transition from the Oligocene to Miocene in the Northern Hemisphere," International Program for Geological Correlation (IPGC), in order to define the correct position of the Paleogene–Neogene boundary and to clarify the problem of age in the brown-coal deposits of Primor'e.

CONCLUSION

(1) In Primor'e, the upper Oligocene–lower Miocene boundary beds are intermediate between the Ust'-Davydovka Formation and the tephritoid bed.

(2) We suggest identifying them as a local stratigraphic unit, the *Fagus*-free Beds, taking into account the specific character of plant remains (the nearly com-

plete absence of beech micro- and macrofossils), and then to classify the unit at the formation or horizon rank.

(3) The position of the Paleogene–Neogene boundary inside the *Fagus*-free Beds still remains ambiguous and needs clarification.

(4) The temperate character of the *Fagus*-free flora and its position in the Tertiary rock sequence of Primor'e provide the basis for correlating it with floras in other North Pacific regions.

(5) We suggest that floras lacking *Fagus* remains originated under paleoclimatic peculiarities of the corresponding epoch that were caused by the influence of the Central Asia anticyclone.

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Planktonic Foraminifers and Pliocene–Quaternary Stratigraphy of North Atlantic Sediments

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Recieved May 18, 1993

Abstract – A detailed study of North Atlantic planktonic foraminifers revealed that most warm-water zonal taxa occur in the Pliocene and Quaternary deposits of the temperate belt. This makes it possible to use the low-latitude stratigraphic zonation of Bolli and Saunders (1985) up to approximately 60°N. In the Pliocene and Quaternary deposits of the Rockall Plateau (Sites 403 - 406, DSDP Leg 48), the following units were established: the *Globorotalia margaritae margaritae*, *Globorotalia margaritae evoluta*, and undivided *Globorotalia miocenica–Globorotalia tosaensis* zones, with the *Globorotalia crassaformis viola*, *Globorotalia crassaformis hessi*, *Globigerina calida calida*, and *Globigerina bermudezi* subzones of the *Globorotalia truncatulinoides* Zone. The use of the low-latitude zonation scheme in boreal regions enables direct correlations throughout most of the Atlantic (with regard to synchronous or diachronous datum levels of zonal species) and makes the development of particular stratigraphic scales for each climatic belt unnecessary. To the north of 60° N, planktonic foraminifers define the Miocene–Pliocene and Pliocene–Quaternary boundaries.

INTRODUCTION

Planktonic foraminifers are known to be very useful in the stratigraphic subdivision of Cenozoic deposits in tropical and subtropical belts. Beginning in the late 60s, several stratigraphic zonation schemes based on evolutionary changes of planktonic foraminifers were developed (Bandy, 1964; Bolli, 1966; Blow, 1969; Berggren, 1972; Lamb and Beard, 1972; Bolli and Premoli Silva, 1973; and others). The scale of Bolli and Premoli Silva (1973), somewhat modified later (Bolli and Saunders, 1985), and that of Blow (1969), are most commonly used to subdivide the low-latitude Pliocene and Quaternary sediments.

Bolli and Premoli Silva developed their scale based on the Caribbean Pliocene–Quaternary deposits, though it was subsequently applied in different oceanic regions. The scale includes the following Pliocene zones: *Globorotalia margaritae* (with the *Globorotalia margaritae margaritae* and *Globorotalia margaritae evoluta* subzones), *Globorotalia miocenica* (with the *Globigerinoides trilobus fistulosus* and *Globorotalia exilis* subzones), and *Globorotalia tosaensis*. The Quaternary corresponds to the *Globorotalia truncatulinoides* Zone, which includes the *Globorotalia crassaformis viola*, *Globorotalia crassaformis hessi*, *Globigerina calida calida*, *Globigerina bermudezi*, and *Globorotalia fimbriata* subzones.

Bolli and Premoli Silva place the Pliocene–Quaternary boundary at the base of the *Globorotalia truncatulinoides* Zone, and this concept is almost universally accepted. The position of the Miocene–Pliocene boundary is a more intricate problem. Sediments with the typical microfauna of the Miocene, *Globorotalia plesiotumida*, and Pliocene *Globorotalia margaritae evoluta* zones are separated by the *Globorotalia margaritae margaritae* Zone (or Subzone) after Bolli (= the *Globorotalia tumida* Zone after Blow), which has a transitional planktonic foraminifer assemblage, in which Pliocene forms appeared against a background of dominant Miocene foraminifers. Bolli and Blow place the Miocene–Pliocene boundary at the base of the *Globorotalia margaritae margaritae* Zone, though moving it to the top of this zone is not inconceivable, above which the Pliocene planktonic foraminifers were represented in all diversity (Krasheninnikov, 1982). Omitting all the discrepancies related to this problem, we accept the lower boundary of the *Globorotalia margaritae margaritae* Zone as the base of the Pliocene.

The tropical Caribbean stratigraphic zonation of the Pliocene and Quaternary was applied to the Atlantic Ocean deposits up to the latitude of Deep Sea Drilling Project (DSDP) Site 370 (32°50'N) near the Moroccan coast. Northward, the taxonomic composition of planktonic foraminifers and other calcareous plankton becomes impoverished, and this fauna gradually loses its stratigraphic importance. Instead, the siliceous plankton, i.e., radiolarians, diatoms, and silicoflagellates, are usually used in detailed stratigraphic subdivision.

This biogeographic situation, a result of climatic zonation, has been adequately addressed in the literature. However, the behavior of tropical zonal index species in subtropical (35° - 45°N), boreal (45° - 56°N), and subarctic (62° - 76°N) regions, as well as the behavior of first and last occurrence datums or datum levels of planktonic foraminifers still remains unclearly understood in many respects.

This work concerns the biostratigraphic subdivision of upper Cenozoic oceanic deposits from the northern-

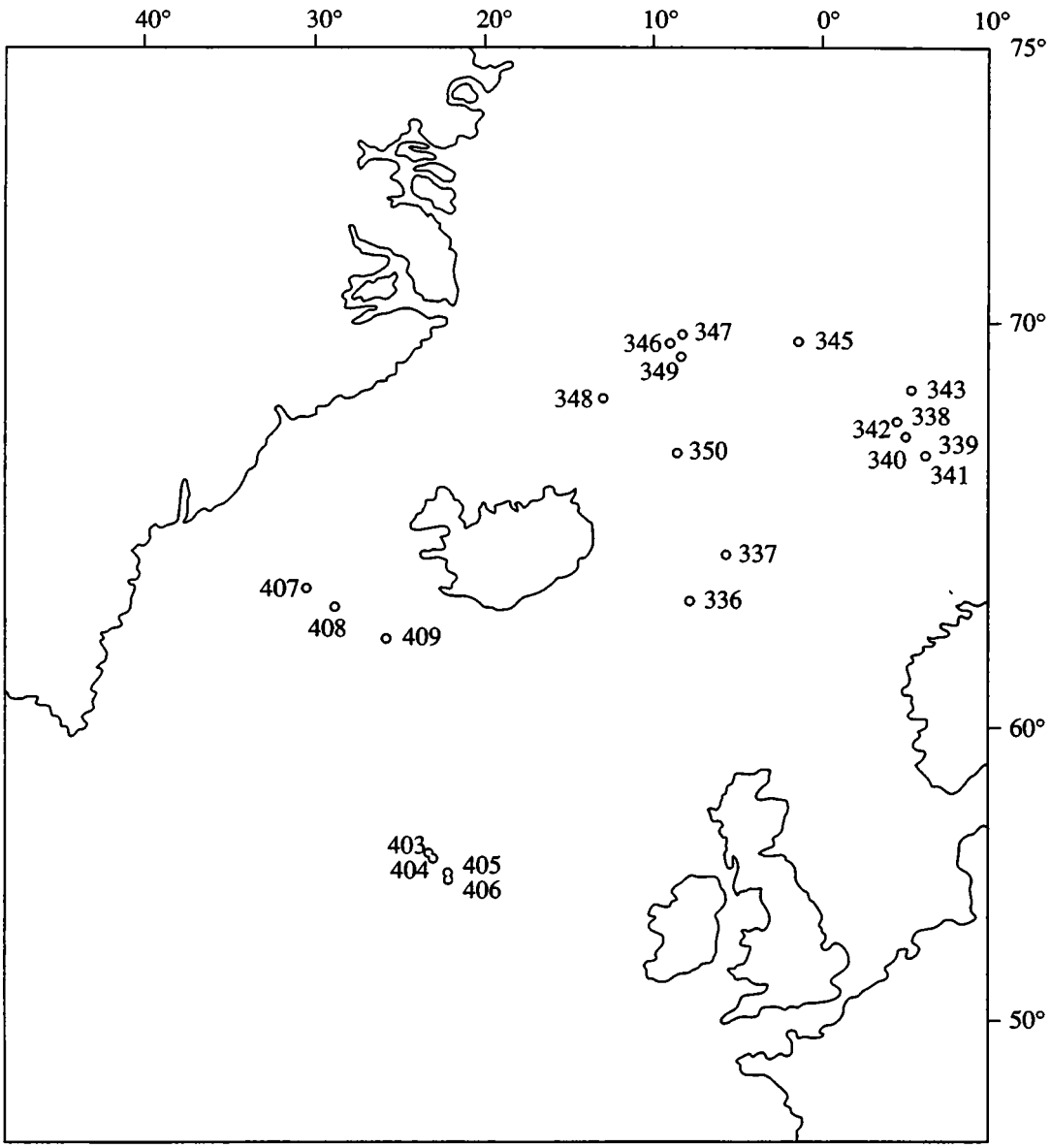

Fig. 1. Location of sites drilled by Legs 38, 48, and 49.

most Atlantic regions (55° - 76°N) by means of planktonic foraminifers from the DSDP samples. Studied samples were collected from the following Sites: 336 - 350, Leg 38, Norwegian–Greenland basin; 407 - 409, Leg 49, southwestward from Iceland; and 403 - 406, Leg 48, Hatton–Rockall Plateau (Fig. 1).

We gratefully acknowledge the DSDP leadership for their presentation of the core material.

HATTON-ROCKALL PLATEAU

In the temperate belt, the Pliocene–Quaternary deposits were recovered at Sites 403 - 406 on the Rockall Plateau, 55° - 56°N (Table 1). The deposits yield an abundance of well-preserved foraminifers with dominant boreal species of a wide stratigraphic range. Due to the dominant *Globigerina bulloides, Globigerinita glutinata, Neogloboquadrina pachyderma, Globorotalia scitula, G. crassaformis,* and *G. inflata* forms, the first biostratigraphic schemes of high-latitude regions were based solely on these species and were significantly less detailed than those of the tropical belt. For example, in Berggren's zonation (1972), the Pliocene–Quaternary sequence on the Rockall Plateau was subdivided into three units, i.e., the *Globorotalia puncticulata, Globorotalia crassaformis*, and *Globorotalia inflata* zones. These units have a large volume, and their exact correlation with Pliocene zones of the tropical scale is difficult.

However, a more recent study of Rockall Plateau deposits allowed us to distinguish the tropical Pliocene *Globorotalia margaritae margaritae, Globorotalia margaritae evoluta*, and *Globorotalia miocenica–Globorotalia tosaensis* zones (Krasheninnikov, 1979) corresponding to the Quaternary subzones of Bolli and Premoli Silva's low-latitude zonation (Bylinskaya, 1991).

The use of the tropical zonation scale in this high-latitude Atlantic region was possible due to the presence of certain Pliocene–Quaternary zonal taxa (*Globorotalia margaritae, G. crassaformis*, and others) in the sediments. These forms, evidently, have a subtropical origin and are cosmopolites in a large part of the Atlantic, though in northern regions they are more scarce.

Let us consider more thoroughly the late Cenozoic biostratigraphy of the Rockall Plateau according to planktonic foraminifers, using the Site 403 as an example (Fig. 2).

Upper Miocene deposits of the *Globorotalia plesiotumida* Zone (= *Globorotalia humerosa* Zone after Bolli) that yield, along with other forms, *Globorotalia conomiozea, G. miozea conoidea* and the index species were identified in Cores 22 - 16.

In Core 14, the appearance of *Globorotalia margaritae margaritae* marks the base of the corresponding zone. Deposits of the zone yield the index species *Globorotalia merotumida* and *G. obesa*, which existed in the Miocene and disappeared in the top of the zone. Such forms as *G. acostaensis* and *G. humerosa*, as well as the sinistral *Neogloboquadrina atlantica. Globigerina nepenthes* and *Globigerinoides obliquus extremus* occurred sporadically. The characteristic Pliocene species *Globorotalia puncticulata* appeared in the middle of Core 13 (Sample 403-13-3, 123 - 125 cm).

All these taxa occur against a background of abundant *Globigerina bulloides, Globigerinita glutinata, Neogloboquadrina pachyderma* dex, and somewhat scarcer *Globorotalia scitula* and *Globigerina quinqueloba*. This dominant group of taxa occurred throughout the Pliocene–Quaternary interval.

In the first half of the Pliocene, this group also included the thermophilic *Orbulina universa*. Later, especially in the Quaternary, it became less common.

At the top of the *Globorotalia margaritae margaritae* Zone, planktonic foraminifers indicate the first significant cooling of the Pliocene climate. It is distinctly expressed in the impoverished foraminifer composition. There were few typical Pliocene species among dominant boreal taxa. Sinistral *Neogloboquadrina pachyderma* appeared here for the first time in considerable amounts (Sample 403-13-1, 54 - 57 cm). The sediments corresponding to the cooling event yield gravel and even small pebbles, probably of an ice-rafted origin.

The occurrence of *Globorotalia margaritae evoluta* in Sample 403-12-6, 106 - 108 cm, marks the base of the corresponding zone with the richest and most thermophilic plankton assemblages of the entire Pliocene–Quaternary interval. The deposits of the zone contain *Globorotalia plesiotumida, G. pseudomiocenica, G. margaritae margaritae, G. margaritae evoluta, Globoquadrina altispira, Globigerina nepenthes, G. venezuelana*, and *Sphaeroidinellopsis seminulina*, forms that became extinct within this unit. There are numerous *Neogloboquadrina atlantica* and *Globorotalia puncticulata* with common *Globigerina decoraperta, G. foliata, Globorotalia acostaensis*, and *G. humerosa* in the assemblage. Turbidite beds with a redeposited late Miocene microfauna (*Globorotalia conomiozea, G. miozea conoidea*) are related to this interval.

Table 1. Data on Leg 48 sites

Hole	Coordinates		Water depth, m
403	56°08.31′N	23°17.64′W	2301
404	56°03.13′N	23°14.95′W	2306
405	55°20.18′N	22°03.49′W	2958
406	55°15.50′N	22°05.41′W	2911

In the lower part of the zone, *Globorotalia crassaformis* with subspecies extremely important for the subdivision of Quaternary deposits first appeared. The sediments of the *Globorotalia margaritae evoluta* Zone yielded *Globorotalia crassaformis crassaformis* and *Globorotalia crassaformis ronda.*

The foraminifer fauna indicates several warm and cool episodes within the described zone. The most essential warming (from Sample 403-12-4, 80 - 83 cm to 403-12-1, 72 - 74 cm) is marked by the presence of tropical *Globigerinoides trilobus, G. sacculifer, G. ruber*, and *Globorotalia tumida*, as well as by an increased amount of *Globigerinella siphonifera* and *Orbulina universa*. The amounts of *Sphaeroidinellopsis seminulina, Globorotalia pseudomiocenica*, and *G. plesiotumida* are also higher here. In the middle of Core 11 (Sample 403-11-4, 90 - 92 cm), a cool event was recorded preceding the warm one, during which the abundance of relatively cryophilic *Neogloboquadrina atlantica* and *Globorotalia puncticulata* decreased (Samples 403-11-2, 53 - 55 cm and 11-1, 59 - 61 cm). Finally, at the top of the *Globorotalia margaritae evoluta* zone, we recorded fauna that indicated a new significant cooling. This climatic event evidently corresponds to the cooling period that occurred 3.0 - 3.2 Ma, which some authors attributed to the onset of glaciation in the Northern Hemisphere (Poore, 1979). The cooling occurred in two steps separated by a short, relatively warm interval, and resulted in the accumulation of small juvenile impoverished fauna with abundant *Neogloboquadrina pachyderma* sin.

Upper Pliocene deposits of Site 403 correspond to the *Globorotalia miocenica* and *Globorotalia tosaensis* zones, which cannot be distinguished in the Rockall Plateau sediments.

A considerable change in foraminifer assemblages occurred at the base of this unit. A group of species, i.e., *Globorotalia plesiotumida, G. pseudomiocenica, Globoquadrina altispira*, and *Globigerina nepenthes*, did not occur in this interval. Only several samples of sediments from the base of the undivided *Globorotalia miocenica–Globorotalia tosaensis* Zone yielded rare *Globigerinoides obliquus extremus* and *Sphaeroidinellopsis seminulina*, as well as a few, possibly redeposited *Globorotalia margaritae*. Within this unit, *Globorotalia praehirsuta, G. inflata*, and *G. cras-*

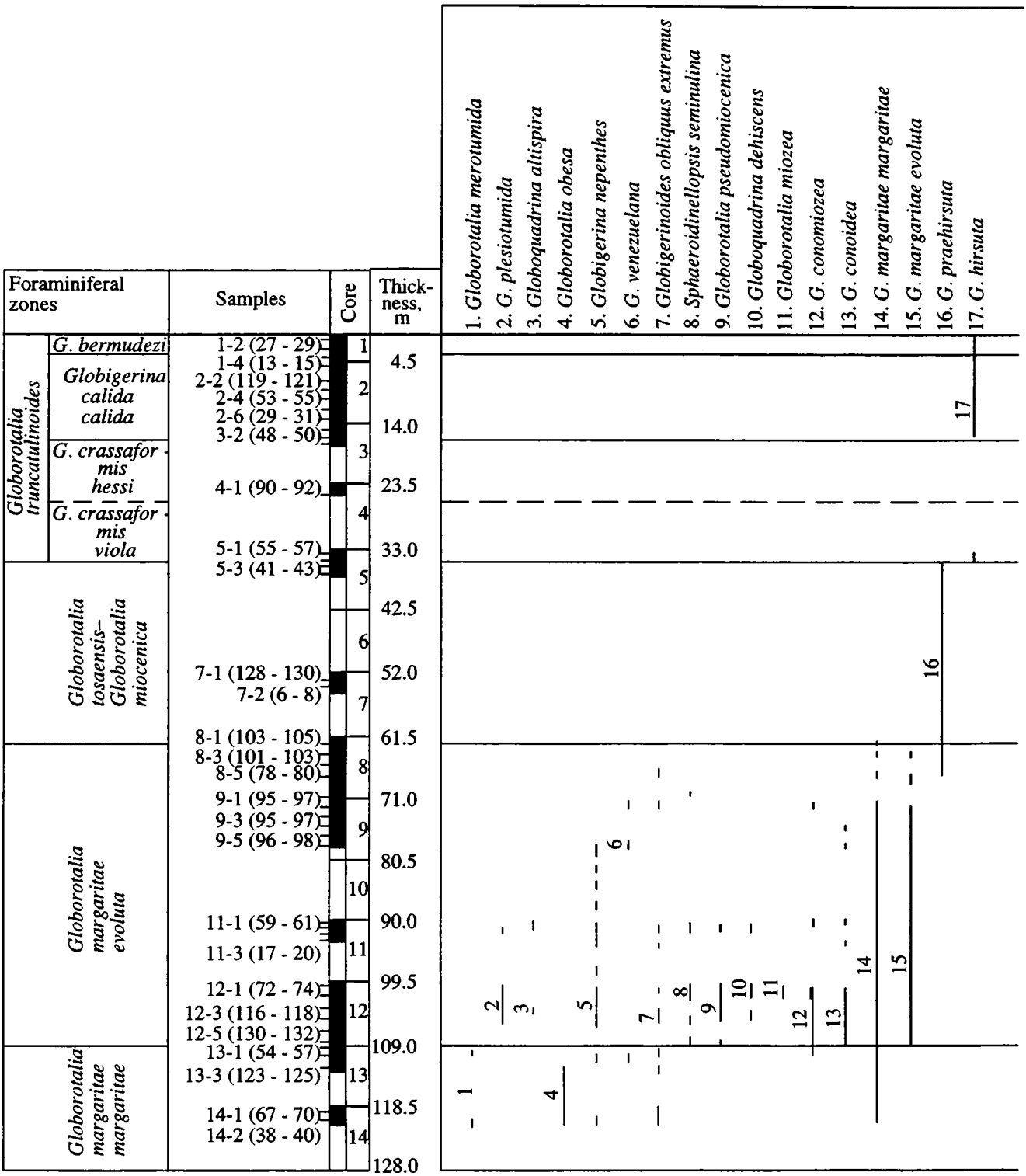

Fig. 2. Stratigraphic distribution of planktonic foraminifers in the Pliocene and Quaternary deposites of Site 403.

Sampled intervals: *Globigerina bermudezi* Subzone - samples 1-2, 27 - 29 cm; 1-3, 57 - 59 cm; *Globigerina calida calida* Subzone – samples 1-4, 13 - 15 cm; 2-1, 24 - 26 cm; 2-2, 119 - 121 cm; 2-3, 20 - 22 cm; 2-4, 53 - 55 cm; 2-5, 30 - 32 cm; 2-6, 29 - 31 cm; 3-1, 28 - 30 cm; 3-2, 48 - 50 cm; *Globorotalia crassaformis hessi* Subzone – samples 3-3, 47 - 49 cm; 4-1, 90 - 92 cm; *G. crassaformis viola* Subzone – samples 5-1, 55 - 57 cm; 5-2, 132 - 134 cm; *Globorotalia tosaensis-Globorotalia miocenica* Zone – samples 5-3, 41 - 43 cm; 5-5, 118 - 120 cm; 7-1, 128 - 130 cm; 7-2, 6 - 8 cm; *Globorotalia margaritae evoluta* Subzone – samples 8-1, 103 - 105 cm; 8-2, 102-104; 8-3, 101 - 103 cm; 8-4, 102-104; 8-5, 78 - 80 cm; 8-6, 101 - 103 cm; 9-1, 95 - 97 cm; 9-2, 95 - 97 cm; 9-3, 95 - 97 cm; 9-4, 78 - 80 cm; 9-5, 96 - 98 cm; 11-1, 59 - 61 cm; 11-2, 53 - 55 cm; 11-3, 17 - 20 cm; 11-4, 90 - 92 cm; 12-1, 72 - 74 cm; 12-2, 8 - 10 cm; 12-3, 116 - 118 cm; 12-4, 80 - 83 cm; 12-5, 130 - 132 cm; 12-6, 106 - 108 cm; *Globorotalia margaritae margaritae* Subzone – samples 13-1, 54 - 57 cm; 13-2, 27 - 29 cm; 13-3, 123 - 125 cm; 14-1, 67 - 70 cm; 14-2, 38 - 40 cm.

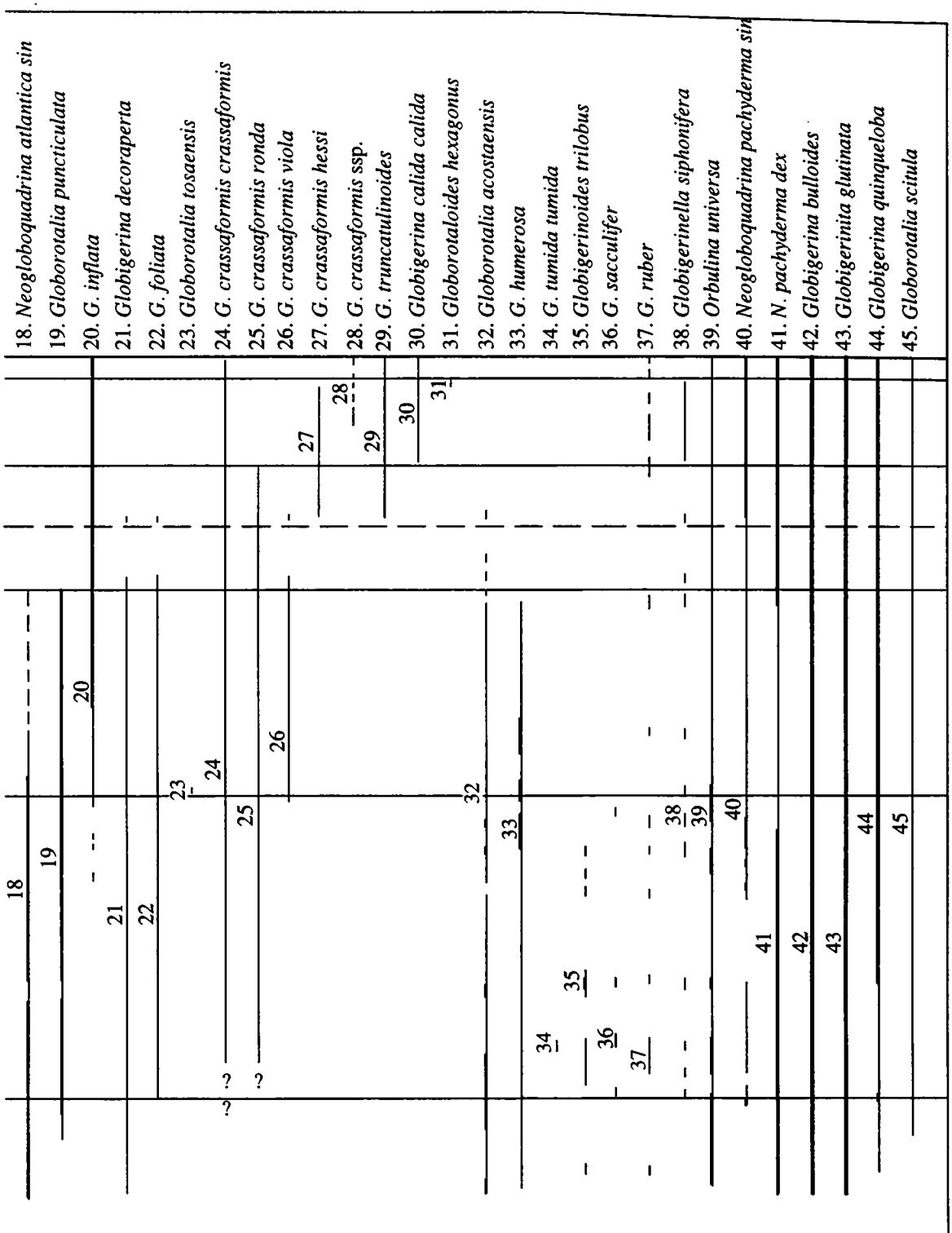
18. Neogloboquadrina atlantica sin
19. Globorotalia puncticulata
20. G. inflata
21. Globigerina decoraperta
22. G. foliata
23. Globorotalia tosaensis
24. G. crassaformis crassaformis
25. G. crassaformis ronda
26. G. crassaformis viola
27. G. crassaformis hessi
28. G. crassaformis ssp.
29. G. truncatulinoides
30. Globigerina calida calida
31. Globorotaloides hexagonus
32. Globorotalia acostaensis
33. G. humerosa
34. G. tumida tumida
35. Globigerinoides trilobus
36. G. sacculifer
37. G. ruber
38. Globigerinella siphonifera
39. Orbulina universa
40. Neogloboquadrina pachyderma sin
41. N. pachyderma dex
42. Globigerina bulloides
43. Globigerinita glutinata
44. Globigerina quinqueloba
45. Globorotalia scitula

Fig. 2. (Contd.)

saformis viola appeared for the first time and progressed rapidly. The *Globorotalia crassaformis* group becomes an important component of the late Pliocene fauna, and, along with *G. inflata*, largely defines its features. *Neogloboquadrina atlantica, Globorotalia acostaensis*, and *G. humerosa* also occur there. As for the index species, *Globorotalia tosaensis* was found in a single sample, whereas *G. miocenica* was absent in high latitudes.

Due to large gaps in sampling of the upper Pliocene deposits, their paleoclimatic interpretation is impossible.

The Pliocene–Quaternary boundary was defined in Core 5 according to the last occurrence of *Neogloboquadrina atlantica* and *Discoaster brouweri* (nannoplankton) in the Sample 403-5-3, 45 - 46 cm (Müller, 1979). This boundary is marked by the last substantial change in the foraminifer taxa. At this level, or close to it, the Miocene–Pliocene species *Globorotalia praehirsuta, G. puncticulata, G. acostaensis, G. humerosa, Neogloboquadrina atlantica, Globigerina decorapperta*, and *G. foliata* disappeared. Specimens of *Globorotalia hirsuta* occurred commonly, though in the first half of the Quaternary the were very scarce. The *Globorotalia truncatulinoides* species was absent in the lower part of the Quaternary at this latitude.

The *Globorotalia crassaformis viola* Subzone of the lowest Quaternary is defined by the occurrence of the index form, which dominates in the *G. crassaformis* group of this unit. We place the upper boundary of the subzone in Core 4 (Sample 403-4-1, 90 - 92 cm), where the *Globorotalia crassaformis hessi* appeared along with single specimens of the index subspecies. At the base of the Quaternary we distinguished the horizon, dominated dextral species of *Neogloboquadrina pachyderma* but barren of sinistral forms. The same horizon is discussed below (Sites 407 - 409, southwest of Iceland).

The *Globorotalia crassaformis hessi* Subzone is characterized by the extinction of *G. crassaformis viola* and by the presence of numerous *G. crassaformis hessi, G. crassaformis ronda*, and *G. crassaformis crassaformis*. This plexus evidently attains its most intensive flourishing here. In this subzone of the Rockall Plateau deposits *G. truncatulinoides* forms were found (Sites 403 and 406). The extrapolated age of its first occurrence here is approximately 1.25 Ma.

The appearance of *Globigerina calida calida* marks the base of the corresponding subzone. In its lower part, the *Globorotalia crassaformis ronda* disappeared, whereas another peculiar subspecies of the *G. crassaformis* group was found. The latter was previously recorded in the Indian Ocean (Rögl, 1974) and in the tropical Atlantic (Pflaumann and Krasheninnikov, 1978).

The first occurrence level of *Globigerina calida calida* in the Rockall Plateau sequence, when correlated with datum levels of nannoplankton zonal species, approximately corresponds to the Matuyama–Brunhes paleomagnetic reversal, as it is in the tropical Atlantic sequences (Bylinskaya and Golovina, 1990). In the sediments of the Rockall Plateau, *G. calida calida* forms are rather scarce, sometimes single, but occur consistently, making it possible to distinguish the corresponding subzone. Thus, the biostratigraphic scale of low latitudes (Bolli and Premoli Silva, 1973) is also valid northwards.

The top of the *Globigerina calida calida* Subzone is defined by the last occurrence of *Globorotalia crassaformis hessi*, approximately at the level of the last Interglacial, about 0.1 Ma.

The deposits of the *Globigerina bermudezi* Subzone are characterized by scarce specimens of *Globorotalia crassaformis* or by their absence. The species was conceivably expelled from the biocoenosis by *G. hirsuta*, which had similar ecologic parameters.

We were not able to determine the *Globorotalia fimbriata* Subzone (Holocene) in the Rockall Plateau sequence, as its index species occurs only in tropical regions.

Table 2. Data on Leg 49 sites

Hole	Coordinates		Water depth, m
407	63°56.32′N	30°34.56′W	2472
408	63°22.63′N	28°54.71′W	1624
409	62°36.98′N	25°57.17′W	832

NORTH ATLANTIC SOUTHWEST OF ICELAND

Deposits of Sites 407, 408, and 409 (62° - 63°N, Table 2) yielded impoverished planktonic foraminifer assemblages with only 7 - 12 species per sample.

At the top of the Miocene, the *Neogloboquadrina atlantica* species changed their coiling from dextral to sinistral. This event is clearly recorded at Site 407 between Cores 17 and 16, and at Site 408 between Cores 14 and 13. The Pliocene foraminifer assemblage contains an abundant amount of *N. atlantica* species, as well as common specimens of *Globorotalia acostaensis* and *G. humerosa* (Figs. 3 and 4). The main species, which passed into the Quaternary, were *Globigerinita glutinata, Globigerina bulloides, G. quinqueloba, Neogloboquadrina pachyderma*, and *Globorotalia scitula*. The thermophilic *Orbulina universa* is also common there.

At Site 407 (Sample 16-3, 122 - 127 cm), a single specimen of *Globorotalia margaritae* s.l. was found, but it is the only finding of the early Pliocene zonal species in these latitudes.

The second half of the Pliocene is marked here by the appearance of *Globorotalia puncticulata* (Samples 407-11-1, 120 - 125 cm and 408-10-3, 120 - 124 cm) and, somewhat later, by an occurrence of *G. crassaformis crassaformis* (Sample 408-8-1, 100 - 102 cm), though at Site 407, the latter appeared simultaneously with *G. puncticulata*.

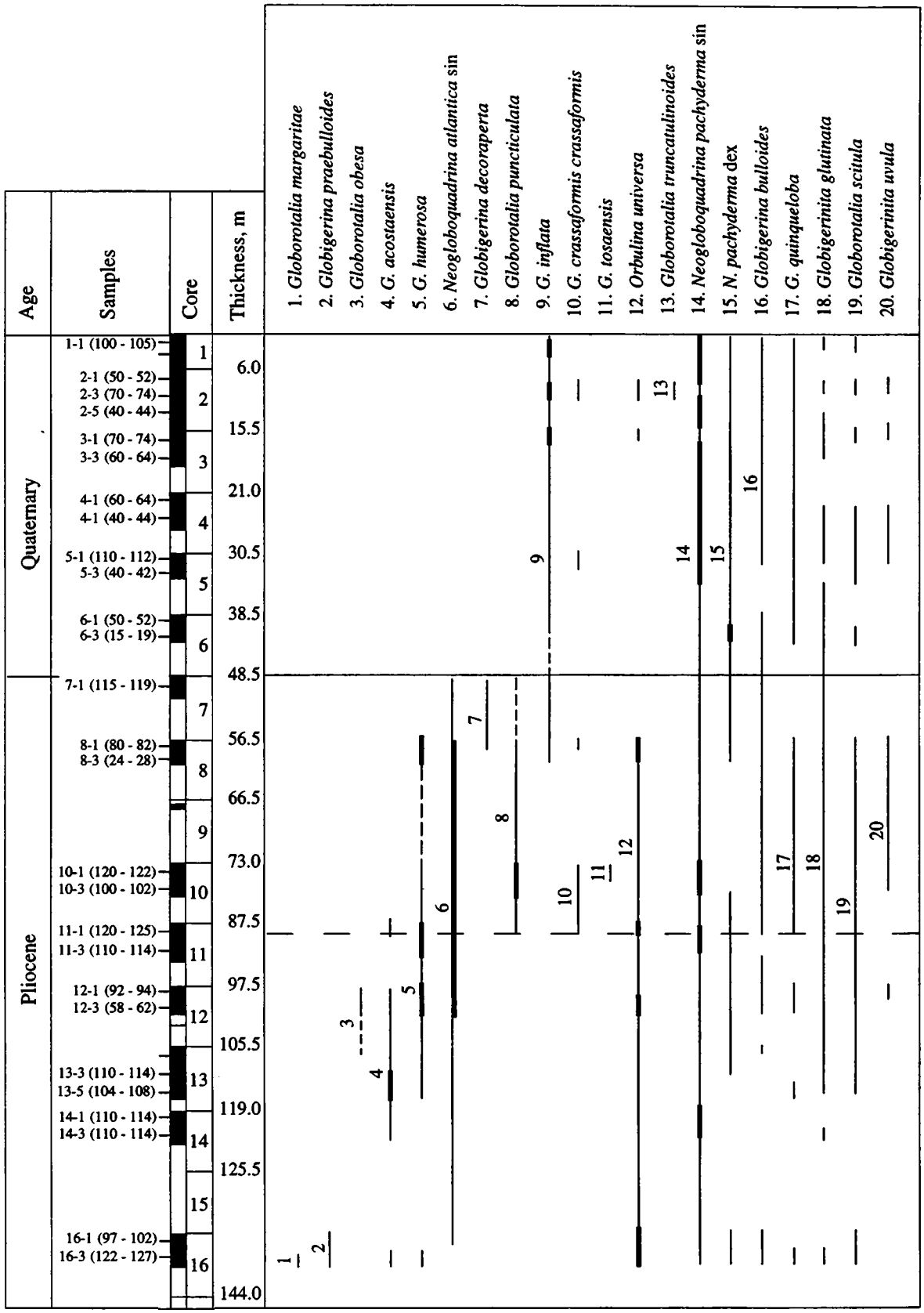

Fig. 3. Stratigraphic distribution of planktonic foraminifers in the Pliocene and Quaternary deposits of Site 407.

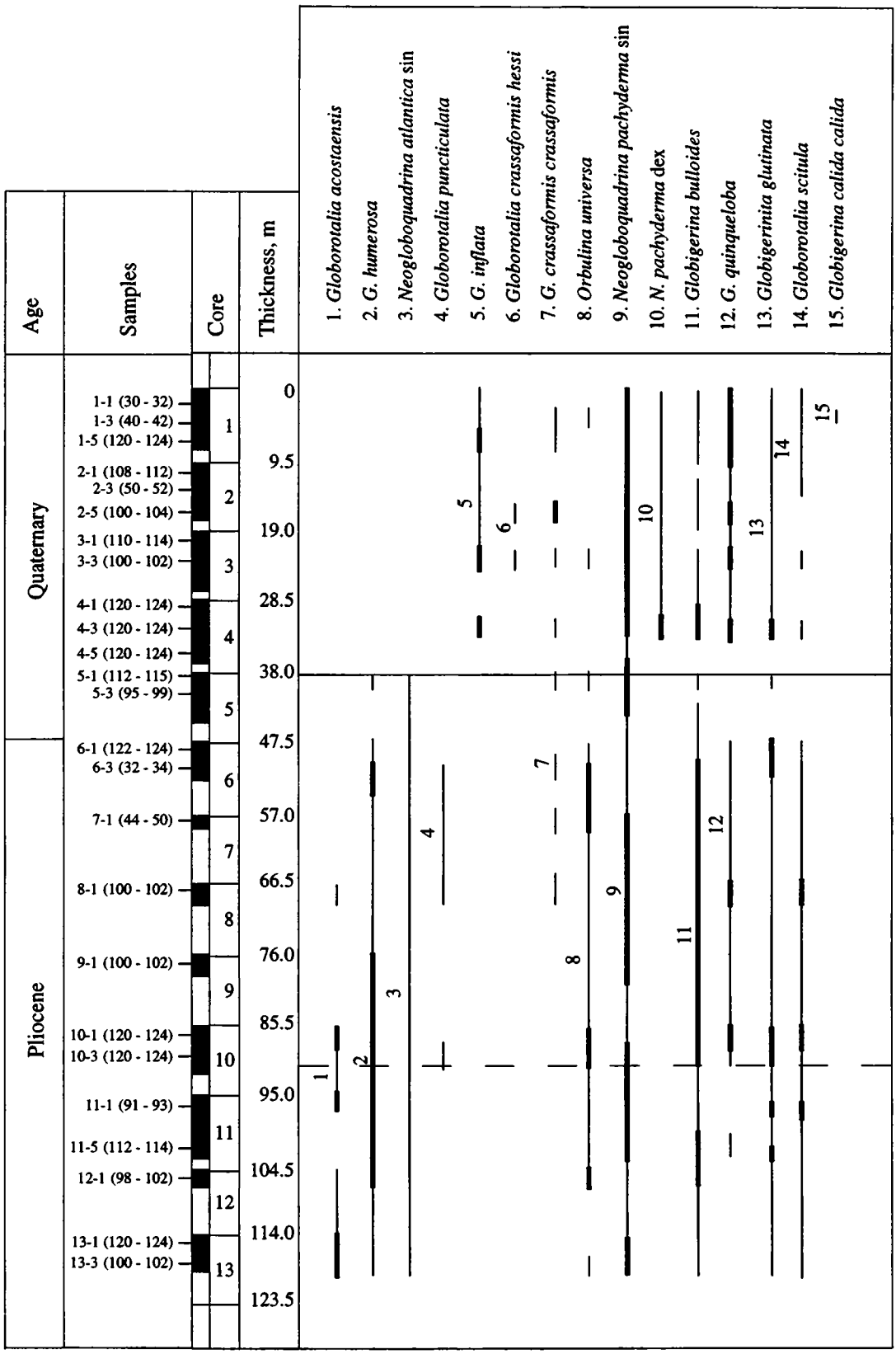

Fig. 4. Stratigraphic distribution of planktonic foraminifers in the Pliocene and Quaternary deposits of Site 408.

Sample 407-10-1, 120 - 122 cm, yielded a single specimen of *Globorotalia tosaensis*, an index species of the corresponding late Pliocene zone. *G. inflata* was found in the upper part of the Pliocene interval (Sample 407-8-3, 24 - 28 cm). The top of the Pliocene (Samples 407-7-1, 115 - 119 cm and 408-5-1, 112 - 115 cm) is the last occurrence level of *Neogloboquadrina atlantica*. At both sites this event coincided with the disappearance of *Discoaster brouweri*. Therefore, the extinction of *N. atlantica* may be considered a good marker of the Pliocene upper boundary.

The Quaternary foraminifer assemblages are more impoverished and cold-resistant than those of the Pliocene. The solitary specimens of *Orbulina universa* were found only in several samples (407-3-1, 70 - 74 cm; 407-2-3, 70 - 74 cm; 408-3-3, 100 - 102 cm; 408-1-3, 40 - 42 cm; and 409-2-1, 50 - 52 cm).

In all three sites close to the base of the Quaternary, we found a horizon with numerous dextral *Neogloboquadrina pachyderma* species (Samples 407-6-3, 15 - 19 cm; 408-4-3, 120 - 124 cm and 409-7-4, 98 - 100 cm). Site 409 is on the crest of the Reykjanes Ridge, and the underlying basalts (Sample 409-7-6) correspond to the Matuyama reversed epoch, i.e., anomaly 2' of about 2.3. Ma. This roughly indicates the age of the horizon (about 1.8 - 2.0 Ma). Because throughout the Pliocene–Quaternary sequence it is a single bed with a pronounced predominance of dextral *N. pachyderma*, it may serve as a regional reference unit, also being established on the Rockall Plateau.

Sites 407 - 409 yielded no barren samples, though many of them, along with impoverished assemblages, contain plenty of shell fragments, the result of strong carbonate dissolution. Levels of dissolution associated with large coolings were recorded at all sites. The most drastic climatic changes in the Pliocene and Quaternary were related to bottom current activity that resulted in sediment erosion and stratigraphic hiatuses. Because planktonic foraminifers together with other calcareous organisms are of little stratigraphic significance in North Atlantic regions, it is impossible to distinguish all the hiatuses if they are unmarked lithologically. We can reveal only the largest unconformities. A large hiatus was distinguished at the Pliocene–Pleistocene boundary at Sites 407 and 408. It is evidently a regional unconformity, which resulted from bottom currents due to one or several late Cenozoic climatic events (Poore, 1979).

Therefore, southwest of Iceland the planktonic foraminifer fauna allows us to define the Miocene–Pliocene and Pliocene–Quaternary boundaries. The Pliocene is subdivided into two parts according to the first occurrence of *Globorotalia puncticulata*. This is, most likely, a migratory event for such a remote northern region. It is necessary to establish the exact age of this event, as it is distinctly featured in the section and may have stratigraphic significance.

Quaternary deposits cannot be subdivided into subzones using planktonic foraminifers. In summary, it is significant that at Sites 407 - 409 the Pliocene–Quaternary horizons of warm-water origin yielded single zonal forms of the Pliocene (*Globorotalia margaritae, G. tosaensis*) and Pleistocene (*Globorotalia truncatulinoides, G. crassaformis hessi, Globigerina calida calida*). These forms indicate the age of deposits, but do not allow us to define the stratigraphic boundaries.

NORWEGIAN–GREENLAND BASIN

Pliocene–Quaternary planktonic foraminifers from deposits of the northernmost Atlantic were studied at Sites 336 - 350 in the Norwegian–Greenland basin (63° - 76°N, Table 3). Significant impoverishment of the calcareous plankton in the composition of both microfauna and microflora hampers the biostratigraphic subdivision of the sediments. Therefore, a large Pliocene–Pleistocene interval spanning 5.2 Ma was only distinguished in Leg 38. The interval was subdivided into two zones: *Neogloboquadrina atlantica* sin and *N. pachyderma*. The base of the latter was dated as 2.8 Ma (Schrader, Björklund, *et al.*, 1976).

A thorough study of planktonic foraminifers from those sites showed the faunal composition to be very poor. The quantity and preservation of the fauna was greatly influenced by the dissolution, which increased during cool periods. As a result, the holes yielded either rare samples containing abundant thin-shelled easily soluble *Globigerina quinqueloba*, or samples with abundant to rare shells of solution-resistant sinistral *Neogloboquadrina pachyderma*, and, finally, absolutely barren rocks.

The Cenozoic deposits of the Norwegian–Greenland basin are characterized by large hiatuses. At some sites, the Quaternary or Pliocene–Quaternary deposits overlie the Eocene (Sites 339, 340, and 347) or Oligocene (Sites 336, 337, 345, and 349) sediments. The Paleogene deposits were only dated according to the siliceous plankton.

Table 3. Data on Leg 38 Sites

Hole	Coordinates		Water depth, m
336	63°21.06′N	07°47.27′W	811
337	64°52.30′N	05°20.51′W	2631
338	67°47.11′N	05°23.26′E	1297
339	67°12.65′N	06°17.05′E	1262
340	67°12.47′N	06°18.38′E	1206
341	67°20.10′N	06°06.64′E	1439
342	67°57.04′N	04°56.02′E	1316
343	68°42.91′N	05°45.73′E	3131
344	76°08.98′N	07°52.52′E	2156
345	69°50.23′N	01°14.26′W	3195
346	69°53.35′N	08°41.14′W	732
347	69°52.31′N	08°41.80′W	745
348	68°30.18′N	12°27.72′W	1763
349	69°12.41′N	08°05.80′W	915
350	67°03.34′N	08°17.68′W	1275

The Miocene sediments occur at many sites (338, 341 - 343, 346, 348, and 350), but almost everywhere they contain only siliceous plankton or agglutinated benthonic foraminifers. Planktonic foraminifers were only found at Site 341. Here, the rocks of Cores 28 - 26 yielded few specimens of *Globorotalia acostaensis* dex, *Neogloboquadrina atlantica* dex and *Globigerina* spp. Two former species suggest the Miocene age of deposits, an age also confirmed by radiolarians and silicoflagellates.

The Miocene–Pliocene boundary is usually defined by a change in the coiling of the *Neogloboquadrina atlantica* species. At Site 341, the Pliocene deposits were recorded in Cores 25 - 23 according to the occurrence of sinistral *N. atlantica*. This high latitude Pliocene index species described by Berggren (1972) occurs in the Norwegian–Greenland basin sediments in considerable amounts and is easily identified, which allows us to confidently subdivide the Pliocene and Quaternary deposits of the subarctic region according to planktonic foraminifers. The *N. atlantica* sin representatives were also found in the Pliocene deposits of Sites 337, 344, and 345. Additionally, the Pliocene sediments at these sites yielded *Neogloboquadrina pachyderma* sin and dex, and rare or single *Globigerina bulloides, G. quinqueloba*, and *Globigerinita glutinata*.

Due to the great number of unsampled intervals, we cannot reconstruct the complete scheme of Pliocene and Pleistocene climatic changes. However, we should note that it is difficult to distinguish a considerable climatic difference between these periods in the subarctic region according to planktonic foraminifers. For example, at Site 341 the Pliocene deposits of Core 24 are barren of foraminifers (341-24-5, 57 - 59 cm) or contain single finds of *N. pachyderma* sin (341-24-2, 55 - 57 cm), evidently the result of a carbonate dissolution caused by a cooling. At the same time, we encountered samples with abundant *N. pachyderma* sin, *N. atlantica* sin, and more scarce *Globigerina bulloides* and *Globigerinita glutinata* (Sample 341-25-2, 56 - 58 cm), which correspond to warm periods.

A similar alternation occurs in the Quaternary deposits, defined in Cores 21 - 1 by the absence of *Neogloboquadrina atlantica*. Intervals with the most thermophilic fauna including, along with the above-mentioned species, *Globorotalia inflata* (Samples 341-20-2, 30 - 32 cm; 9-2, 61 - 63 cm; 7-5, 62 - 64 cm; 4-2, 56 - 58 cm) and even *Orbulina universa* (Samples 341-7-2, 33 - 35 cm; 5-2, 94 - 96 cm) alternate with those that bear only rare or single shells of *Neogloboquadrina pachyderma* sin or are even barren (Samples 341-17-2, 40 - 42 cm; 12-5, 40 - 42 cm; 12-2, 50 - 52 cm; 10-2, 50 - 52 cm; 8-5, 63 - 65 cm; 5-5, 64 - 66 cm; 3-2, 65 - 67 cm; 1-2, 72 - 74 cm).

In addition, we did not record any variation in the composition of Pliocene and Quaternary planktonic foraminifers in different parts of the Norwegian–Greenland basin studied in Leg 38. Only at the northernmost Site 344, situated westward from Spitsbergen, is the carbonate fauna poorer then elsewhere. At this site, *N. atlantica sin* occurred in Cores 32 - 14 (Pliocene), whereas most of the Quaternary deposits are devoid of carbonate plankton and yield diatoms and radiolarians. Rare forms of *N. pachyderma* sin were only found in rocks of Cores 5 - 1.

Returning to the problem of hiatuses, we must note that the most evident of them correspond in age to the Pliocene or its part and, possibly, to the base of the Quaternary, as the preserved Pliocene deposits were found only at Sites 337, 341, 344, and 345.

The Ocean Drilling Program (ODP) Leg 104 in the Norwegian Sea led to the conclusion that foraminifers are unsuitable for defining exact levels of the Miocene–Pliocene and Pliocene–Pleistocene boundaries in the region, because the first and last occurrences of sinistral *N. atlantica* was dated as 6.2 and 2.3 Ma, respectively (Spiegler and Jansen, 1989). However, as was stated above, in the adjacent area (southwest of Iceland, only 1° - 2° to the south from the described sites), *N. atlantica* sin disappeared almost simultaneously with *Discoaster brouweri*, and, in any case, the former event did not precede the latter. The extinction of *D. brouweri* is considered to be an isochronous event in the World Ocean, occurring 1.87 Ma (Hills and Thierstein, 1989). Therefore, we can suppose that in the Norwegian Sea *N. atlantica* disappeared a little earlier, because in two sites of Leg 104, horizons with *N. atlantica* sin and *N. pachyderma* sin are separated by a short barren interval.

Accordingly, the study of late Cenozoic stratigraphy of the Norwegian–Greenland basin (northward from 63°N) suggests that the extreme impoverishment of planktonic foraminifers allows the conventional stratigraphic subdivision of sediments only into the Miocene, Pliocene and Quaternary units.

CONCLUSION

(1) Planktonic foraminifer assemblages from the northern boreal Atlantic (Sites 403 - 406, Rockall Plateau) lack a number of species typical of synchronous sediments in the lower latitudes, and certain index species occurred in limited amounts. Nevertheless, the majority of Pliocene and Quaternary zonal taxa established in the low latitudes (Bolli and Premoli Silva, 1973; Bolli and Saunders, 1985) were found in the studied region. The following Pliocene units were identified in the Rockall Plateau deposits: *Globorotalia margaritae margaritae, Globorotalia margaritae evoluta*, and undivided *Globorotalia miocenica–Globorotalia tosaensis* zones. The Quaternary corresponds to the *Globorotalia truncatulinoides* Zone, which includes the *Globorotalia crassaformis viola, Globorotalia crassaformis hessi, Globigerina calida calida,* and *Globigerina bermudezi* Subzones. Only the Holocene *Globorotalia fimbriata* subzone was not distinguished here.

(2) Northwards, in the area transitional between the Boreal and Subarctic regions (Sites 407 - 409, southwest of Iceland), planktonic foraminifer assemblages are

extremely impoverished. Only the Miocene–Pliocene and Pliocene–Quaternary boundaries can be defined in this area. Zones and subzones are not distinguished. The Pliocene is subdivided into two parts according to the appearance of *Globorotalia puncticulata*. However, this is most likely a migratory level, and its age must be refined.

(3) In the Subarctic region (Sites 336 - 350, Norwegian–Greenland basin), planktonic foraminifers allow a merely conventional definition of the Miocene–Pliocene and Pliocene–Quaternary boundaries. The species composition is extremely poor, and the degree of compositional changes at stratigraphic boundaries is minimal. Thus, the boundaries are not reliably established.

(4) The effective use of planktonic foraminifers for the Pliocene and Quaternary stratigraphy is limited by the latitude of the Rockall Plateau (55° - 56°N). Northwards they lose their stratigraphic significance, giving way to siliceous plankton, and can be used only as a supplementary group.

(5) In the case where a certain datum level coincides with several paleontological events, we can more or less reliably establish its synchronism or asynchronism in different parts of the Atlantic (for example, the synchronous disappearance of *Globorotalia crassaformis hessi* and the first occurrence of *Globigerina calida calida*; the diachronous appearance of *Globorotalia truncatulinoides* and *G. puncticulata*). Otherwise, the planktonic foraminifer datum levels must be correlated with the siliceous plankton zonation or with the paleomagnetic scale.

(6) The use of low-latitude zonation in the boreal region allows us to directly correlate sediments throughout a vast area in the Atlantic Ocean and, in our opinion, renders the development of separate stratigraphic scales for each region unnecessary. The correlation of different schemes will always be more difficult than the establishment of the synchronism or diachronism of datum levels within a single scale for different climatic regions.

(7) Planktonic foraminifers are very sensitive to temperature changes. On the Rockall Plateau they indicate warm-water conditions in the early Pliocene and a drastic deterioration in climate during the late Pliocene and Quaternary. Within these intervals there were climatic fluctuations of lesser rank. Therefore, when available, the continuous deep-sea drilling cores allows us to use the climatostratigraphic method for the entire Pliocene and Quaternary intervals. Its stratigraphic potentials are higher than those of biostratigraphy, though the evolutionary datum points of the latter are necessary to form a valid basis for the stratigraphic scale.

The study was supported by the Russian Foundation for Fundamental Research (project no. 93-05-8568).

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